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CHANNEL PATTERNS OF THE UPPER RED DEER RIVER, ALBERTA, CANADA

by

(C) KNUT OLAF NIEMANN

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

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EDMONTON, ALBERTA

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THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled Channel Patterns of the Upper Red Deer River, Alberta, Canada

submitted by Knut Olaf Niemann

in partial fulfilment of the requirements for the degree of Master of Science

ABSTRACT

The Red Deer River, from the Rocky Mountain foothills across the high interior plains of western Alberta, reveals a variety of channel patterns. These encompass single-channel, relatively straight reaches through to multiple-channel, braided segments. This study attempts to specifically identify the various channel patterns and to place these in context with the factors which control their development.

Hydrographs for the gauging station situated near the mouth of Burnt Timber Creek are plotted for the period from 1974 to 1977. These plots show: (1) large variations in the discharge patterns exist on an annual basis and (2) large seasonal fluctuations in flow patterns occur. The seasonal variations reflect melting of the mountain snow pack in the headwaters and occur as floodwaves passing through the system. Yearly variations are determined by the size of the snow pack and the rate of melting.

The surficial geology and geomorphology of the Red Deer River Valley area are discussed in some detail. The present Red Deer River flows through a preglacial valley in which large quantities of coarse clastic material were deposited during deglaciation. These gravels have subsequently been reworked by the Red Deer River and distributed within the terraces and floodplain. The bank materials are, therefore, composed predominantly of gravel. There also occur a number of bedrock outcrops and along the banks in the study reach a minor amount of lacustrine silt and clay. The channel sediments show a well defined

negative correlation of size against distance. A break occurs in the grain size trend at the foothills-plains boundary, with the foothills section having a greater rate of decrease in the size of the material than is found in the downstream portions. No significant relationship is shown between the sediment texture, degree of sorting and longitudinal profile slope.

Differentiation of channel patterns is accomplished through a width ratio (Wr) devised for this study. This ratio differentiates between single-channel and multiple-channel reaches. Comparison of the values obtained using the Wr values and the longitudinal profile indicates a weak negative relationship. Division of the longitudinal profile into six, distinct, secondary concavities reveals a stronger negative relationship between the channel slope and channel pattern. A positive relation between size and sorting of the channel sediment and the slope exist for most areas within these secondary concavities. This trend does not occur in two situations where major tributaries join the Red Deer River. The gentler sloped segments are therefore associated with higher Wr values. The occurrence of bedrock outcrops at the downstream limits of the concavities is also interpreted as being instrumental in increasing the Wr values. The outcrops act as controls which partially restrict flow during the higher discharge events. This leads to the accentuated deposition of alluvium upstream from the controls, causing channel division and enhanced local bank erosion.

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CHAPTER ONE

INTRODUCTION

1.1 STUDY AREA AND PROBLEM

The main objective of this study was to identify and explain the variation of channel patterns found within a reach of the Upper Red Deer River, Alberta. A second objective was to gather base line data for River Engineering, Alberta Environment, for their bank stabilization studies in this portion of the Red Deer River. The study area occupies a zone across part of the Rocky Mountain Foothills and the high interior plains (Figure 1.1). The main local settlement is Sundre (population 1,000) which lies approximately in the middle of the study reach. This segment of the river extends approximately 75 kilometres from the confluence with William's Creek in the west to the confluence with Raven River in the northeast. For ease of reference, specific locations along the study segment will be referred to by distance (in kilometres) from the upstream end of the reach.

Near the confluence with William's Creek the Red Deer River displays a multiple-channel floodplain but as it passes downstream through a more constricted portion of the valley it reverts to a single-channel and then, further downstream, again regains a multiple-channel form. The latter transition occurs at the eastern boundary of the foothills. More subtle variations of the channel pattern occur over sub-reach distances from straight or slightly sinuous to intensively braided sections.

Major factors contributing to the varieties of channel pattern are based primarily on the nature of the area's hydrology and the past

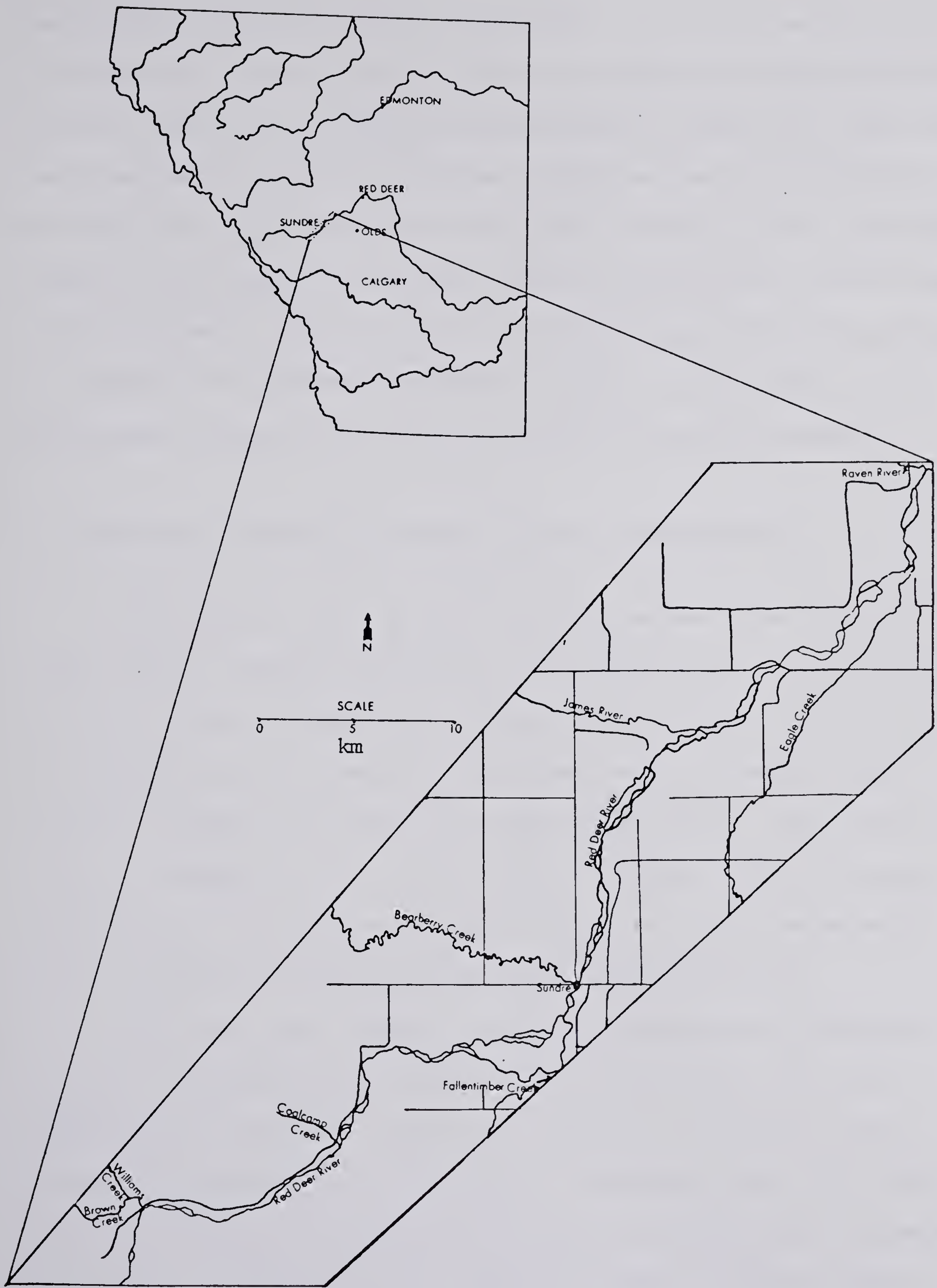


Figure 1.1 Location Map

and present contributing sources of sediment. It was impossible to investigate fully the hydrology of the entire study area because available data are scanty, but pertinent generalisations are presented. More abundant observations relative to sediment supply sources were practicable, and these form a major portion of the thesis. Because of the constraints imposed by one summer of field work the data obtained were largely qualitative in nature. This has limited the scope of some of the discussions. An analysis of contemporary sediment transport, in particular, was impracticable although this factor is probably of extreme importance.

1.2 SELECTED LITERATURE ON FACTORS PROMOTING RIVER BRAIDING.

Although the exact causes of braiding are not known with certainty several, probably important, contributing factors have been cited in the literature. Included are variables such as apparently distinctive characteristics of bank resistance and channel sediments, highly variable discharge, and relatively steep channel slopes. These factors were thus singled out for closest attention in terms of data collection and analysis. It is appropriate here, though, to briefly review some of the pertinent literature dealing with these major factors.

The first major variable related to channel pattern development is that of discharge characteristics. Leopold and Wolman (1957) concluded that the width of a channel is controlled by the bankfull discharge. In reference to the White River, Washington, Fahnestock (1963) found that variations in discharge and the previous flow history appeared to be almost as important as the quantity of water in the development

of a braided channel. He noted that;

> Rapid adjustment to increasing flow and therefore increasing bedload in non-cohesive material appears to take place by increasing width and depth in the following manner. The initial channel widens and may even grow shallower, until at some critical point the critical tractive force is less than that necessary for transport of the bedload. The material deposited causes further widening and a decrease in the tractive force. Deposition or cutting of adjacent channels eventually brings the bar above water, resulting in the development of two channels similar in shape to the original channel but adjusted to the higher flow condition.
> (Fahnestock, 1963, p. 16).

Fahnestock and Bradley (1973) noted that two glacially derived streams, the Knik and Matanuska Rivers, Alaska, have quite distinctive morphological characteristics and behave in response to differing formative discharges. They found that, "nine sets of air photos taken over a 30 year period show that the main elements have stayed approximately the same during that time, although details have changed" (Fahnestock and Bradley, 1973, p. 235). On the other hand "the Matanuska River at normal summer flows is actually changing the details of its complex pattern..." (Fahnestock and Bradley, 1973, p. 237). They contended that the Knik River is adjusted to flows of a magnitude greater than those resulting from annual runoff. In this case the periodic draining of Lake George at the snout of Knik Glacier is the cause of these catastrophic discharges. The Matanuska River did not experience these large scale flows. The annual, large scale, morphological changes were therefore considered to indicate an adjustment to the annual peak flows. One, apparent, significant factor was that the channel slopes differed considerably. The Matanuska River had a slope which was five times that

of the Knik River and, according to the authors, the former experienced a comparatively high and widespread bedload transport. This would have also contributed to the comparative channel instability of the Matanuska River.

Schumm (1977, p. 106) stated that, "although the records are short, they indicate that rivers with high ratios of peak to mean discharge are morphologically different from rivers with low ratios." He noted that this is the only factor which can explain the pattern differences of some rivers. The streams experiencing high ratios tend to be braided whereas those where the differences are less evident are single-channel types. This point was also emphasised by Miall (1977) who noted that braided rivers experience higher flood peakedness, lower total discharge and higher monthly variations in discharge than do meandering streams. He stated that these characteristics of the hydrological regime are essential for the development of braiding processes, and that selective deposition of material may be initiated by variations in discharge.

Edgar's (1973) laboratory experiments produced an excellent correlation between channel width, slope and discharge. He found that over a range of slopes the channel width varied with approximately the square root of discharge. Gesseler (1971) stated that an increase in discharge will increase the depth of flow and therefore increase the erosive potential of a stream. Edgar (1973), on the other hand, found that the channel slope and not the discharge had the greatest influence in determining the depth of a channel.

A hydraulic factor which has not received much attention in the literature is that of the role of seepage from the channels into

the surrounding floodplain material, and the potential effects of this on river competence and floodplain morphologies. Harrison and Clayton (1975) considered that seepage losses of water have the potential of greatly decreasing the competence of a stream. This factor will be discussed further in later sections.

The second major factor in channel pattern development is that of relative bank stability, largely determined by alluvium properties. Fahnestock (1963), in his study of the braided White River, Washington, found that a substantial proportion of the fluvially transported material originated from erosion of the banks. He contended that, "...at the same time and discharge, both braided and meandering reaches were present on the valley train" (Fahnestock, 1963, p. 58). A specific channel pattern was not restricted to a certain location on the valley train, he noted, and changes in the floodplain morphology could not be related solely to variations of bank materials. Therefore, other factors must also be instrumental in promoting variations of channel patterns.

The degree of bank stability was interpreted by Brice (1964) as one of the important controlling factors of channel pattern development on the Loup River, Nebraska. The other main factor was channel slope. Instability was related to the texture of the bank materials but in this area predominant control was considered to be exercised by the riparian vegetation. He noted that;

> Along reaches where vegetal resistance to bank erosion is high, the river course is nearly straight and tends to narrow. Along reaches having low vegetal resistances to bank erosion, as in the lower segment, the river developed a broad, shallow, and somewhat braided course.
> (Brice, 1964, p. 34)

Schumm (1960, 1963) showed that the percentage of silt and clay contained within the bank sediments has a significant effect on the width-depth ratio and channel pattern. Using data from a large number of rivers, he demonstrated that high channel sinuosity and a low width-depth ratio correlated with a high percentage of silt and clay. Brice (1964) pointed out, however, that the stratigraphic distribution of these fines within the banks may be of equal importance to the amount of cohesive material contained. He contended that banks with evenly distributed fine material throughout will behave differently to those where the fines are concentrated in discrete layers.

Coleman (1969), in his study of the Brahmaputra River, India, found that although the sediment load was similar for two reaches, braiding occurred only in areas where the banks were unstable. Such instability was governed by the interaction of numerous factors, including cohesion and variability of bank materials, rate of rise of the river level during the passing of a floodwave, the numbers and positions of major channels, angles at which the thalwegs approached the banklines and the formation of large bedforms.

Church (1972) indicated that highly erodible banks resulted in rapid channel migration. This in turn resulted in shoaling and the distribution of debris within the channel, forming bars and islands, which led to the division of channels around these forms.

Relative bank stability and material characteristics therefore have two main impacts on channel pattern development. The first is the contribution of material to the sediment load of the river. The second is the resistance offered to erosion, partly reflected by the

width-depth ratio of the channel. Highly erodible banks will contribute potentially large amounts of sediment and promote a large width-depth ratio, both of which are important elements of river braiding.

The third major element of channel pattern development is that of sediment availability and related sediment load characteristics. The latter deals more specifically with the relative quantity of material moved as bedload as well as the sorting of the sediment comprising the load. Leopold and Wolman (1957, p. 50) stated that, "braiding is developed by sorting as the stream leaves behind those sizes of the load which it is incompetent to handle." They found that bar evolution was initiated by the deposition of coarsest material, which formed the nucleii of embryo bars. This bar initiation was followed by the deposition of progressively finer material as the competence of the stream decreased. This reduction may result from either a decrease in discharge (and the accompanying reduction in flow velocities) or a decrease in depth of flow over the developing bar surface, thereby lowering the shear stress exerted on the material.

Chien Ning (1961, p. 754) found that braiding on the Yellow River, China, was the result of "the contradiction between the incoming load and the sediment carrying capability of the flow." In other words the excessive load introduced into the Yellow River resulted in deposition and the creation of bars. He did not detail the sedimentary characteristics of the alluvium in terms of sorting.

Fahnestock (1963) noted that on the White River those braided areas which were not restricted in their evolution by bank stability were controlled by the amount of bedload. The nature of the bedload

was not specifically stated. However, the source materials consisted predominantly of till and colluvium, implying a heterogeneous makeup.

A majority of workers have noted the importance of heterogeneous channel sediments. Leopold, et al., (1964), though, found that braiding may occur in homogeneous fine material as well as in the coarse gravels normally cited. They stated that "even in fine material, however, irregular deposition of bars and bank erosion may produce a braided pattern" (Leopold, et al., 1964, p. 292). Fahnestock (1963) attributed this to the occurrence of areas where local competence was exceeded by sediment supply, leading to the deposition of part of the excessive load.

A sometimes erroneous implication arising from these studies is that braided rivers must represent aggrading conditions. Fahnestock (1963) pointed out that a section of the White River experienced net degradation during one field season. He stated that, "the net loss of elevation...suggests that material was derived from both bed and banks and that braiding occurred in a degrading reach" (Fahnestock, 1963, p. 59).

A related factor is that of channel slope. There is some question as to whether this factor is dependent or independent with respect to channel pattern development (Smith, 1973). Disagreement as to the relative importance of channel slope in dictating channel morphology has also arisen. Fahnestock (1963) noted that the channel slope did not explain the variations in the channel patterns he encountered. He found that braided channels developed on a wide range of slopes (from 0.01 to 0.2) and that multiple-channel reaches developed

on slopes both lower and higher than those observed for single-channel reaches. For example, he stated that if the channel slope exceeded a certain value, deposition and the subsequent formation of bars did not occur because of accentuated flow velocities. In other words, there appeared to be an upper threshold limit of slope although the lower limit was not readily apparent. Fahnestock (1969) found that the slope of the Slims River, Alaska, was significantly less than that of the White River, Washington, although braided channel patterns occurred throughout both. Similarly, Fahnestock and Bradley (1973) found slopes of 0.001 and 0.005 for the Matanuska and Knik Rivers respectively. Knighton (1976) noted that, contrary to most findings, slopes were gentler in braided reaches than in single-channel reaches for his study of Nigel Creek, Alberta. He found that braiding occurred in depositional depressions and concluded that bars were formed in zones in which bedload material accumulated. These bars diverted the flow and caused channel divisions.

An opposing view held by some workers is that at a given discharge the slope of a channel dictates the channel morphology. Leopold and Wolman (1957) demonstrated, through observations of natural and laboratory channels, that divided reaches had higher channel slopes than those of single-channel reaches. They recorded a six-fold increase in the slope after channel division on a section of the Green River, Wyoming. From laboratory experiments Schumm and Kahn (1972) observed that meandering thalwegs were formed at lower slopes than for multiple-channel reaches. This confirmed a number of observations made by earlier workers, (Figure 1.2), but failed to take into con-

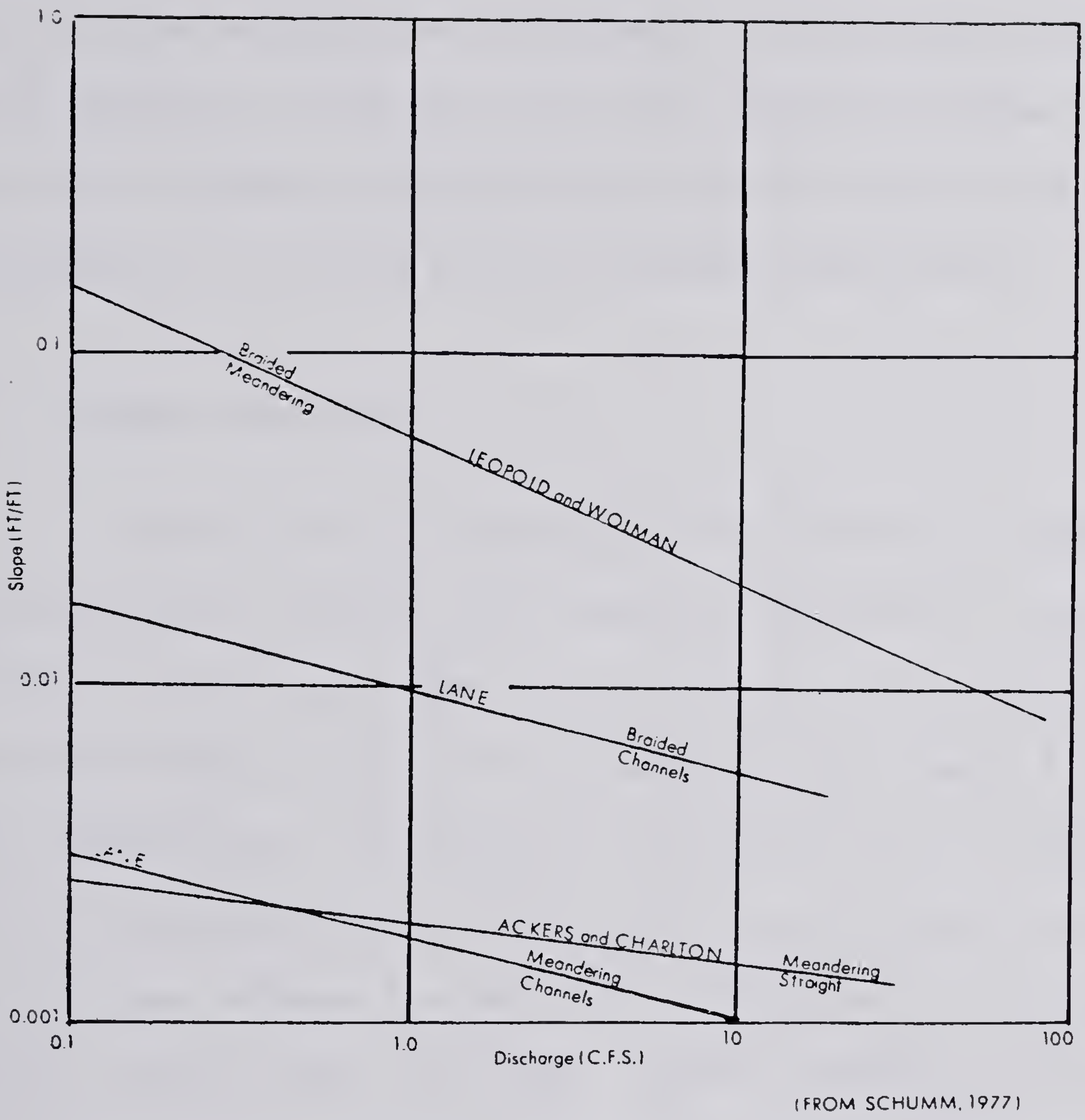


Figure 1.2 The relationship of slope and discharge to channel pattern development.

sideration the possibility suggested by Fahnestock (1963), of an upper limiting slope threshold for braid development. This, also, does not take into consideration the effects of sediment size and sorting on channel pattern development.

Barker (1976) speculates, through numerical analysis, that meandering and braided rivers are a function of both slope and width. If the slope and width are sufficiently low, the river will meander while braiding is favoured with higher slopes and widths. He concludes that both braiding and meandering may occur on equal slopes if the resistance of the banks to erosion varies significantly.

1.3 GENERAL CONCLUSIONS

The main factors contributing to variations of channel pattern development are thus far from completely understood in terms of their relative significance. The studies which have been conducted on distinctive channel patterns are so numerous that a complete review is not attempted here. Some general conclusions may be reached, though, from the previously noted studies. The first is that a majority of workers have stressed the importance of relatively steep channel slopes, highly erodible banks, and highly variable discharge in the development of braided channel patterns. The main criterion, however, appears to be the pronounced development of bedload transport. Fahnestock (1963, p. 61) concluded by stating that;

> Braided channels cannot develop without bedload. The common element in all explanations of braiding appears to be movement of bedload with local deposition within the channel, causing the diversion of flow from one channel into one or more other channels and the development of islands. The rate of pattern change

appears to be directly related to the amount of bedload.

The comparative significance of all the other factors discussed, in relation to bedload transport, is not known. It appears that these relationships may vary subtly from one environment to the next and the mix of factors responsible for river braiding will vary from one environment to the next. A majority of the studies of braided rivers, cited in this review, have been conducted on glaciofluvial valley trains, sandar or outwash plains. For example, Church (1972) described typically rapid sandar aggradation on Baffin Island, Canada, and noted that important conditions favouring sandar development were abundant sediment supply and frequent floods, competent to move the supplied debris. Although these general requirements may pertain to all braided rivers their specific relationships must vary within a quite broad numerical range.

Finally, theoretical conclusions based on flume experiments must be treated with caution. The scaling down of a natural fluvial system inevitably introduces unrealistic elements so that experimental results may not truly reflect natural channel pattern evolution.

CHAPTER TWO

HYDROLOGY, CLIMATE AND VEGETATION

2.1 INTRODUCTION

The hydrological regime of a river has certain important implications for the morphological development of the river, as noted in Chapter One. Church and Gilbert (1975) stated that the characteristics of the flows determine the nature of the sediment transport regime. They found that in a proglacial situation, a highly variable discharge pattern coupled with a large number of moderate flow events tend to cause rapid evolution and destruction of channel forms and initiate changes in the channel pattern. It is clear, however, as pointed out by Gregory and Walling (1973), the relationship is not as simple as that implied above. Rather, the discharge patterns interact with other variables, such as bank materials and vegetation to produce distinctive channel patterns.

Two relevant concepts, in this respect, are those of bankfull discharges and dominant discharges. Harvey (1969), stated that in many instances the bankfull discharge is synonymous with the dominant discharge. The latter is that "...which governs the size and shape of the channel and to which the landform is adjusted" (Harvey, 1969, p. 82). He noted that maximum competence is normally associated with the bankfull limits, when certain relationships between discharge and sediment transport are at an optimum. Greater discharges, above bankfull, will

result in changes in the relationship between the two variables of discharge and sediment transport. Harvey (1969) also stated that streams with strong peak flow events (that is, a highly variable discharge with a number of large flood peaks) are adjusted to bankfull limits while streams which experience low discharge variability are adjusted to flows of greater magnitude than bankfull. In other words, it was implied that the latter are adjusted to flood events with relatively long recurrence intervals.

In this study identification of the bankfull limits was based on two criteria. For the channel cross sections, prominent breaks in the slope between the channel and the floodplain were used (as suggested by Schumm 1960, Leopold et al, 1964 and Speight, 1965). Distinctive vegetation transitions (Speight, 1965) were employed to delineate the bankfull widths for calculation of the width ratios (see Chapter Five). Surfaces consisting of bare gravel, or gravel covered with willow and immature aspen poplar trees, were interpreted to be contained within the bankfull range. Areas with spruce and mature aspen poplar vegetation were designated as floodplain surfaces.

2.2 MAJOR HYDROLOGICAL CHARACTERISTICS OF THE STUDY AREA

Discharge data, for four years, from the Red Deer River gauging station near the confluence of Burnt Timber Creek were used as the base of the following discussion. A station formerly operated as Sundre was discontinued in 1973. However, for that year there was an overlap of discharge data for both the Sundre and Burnt Timber

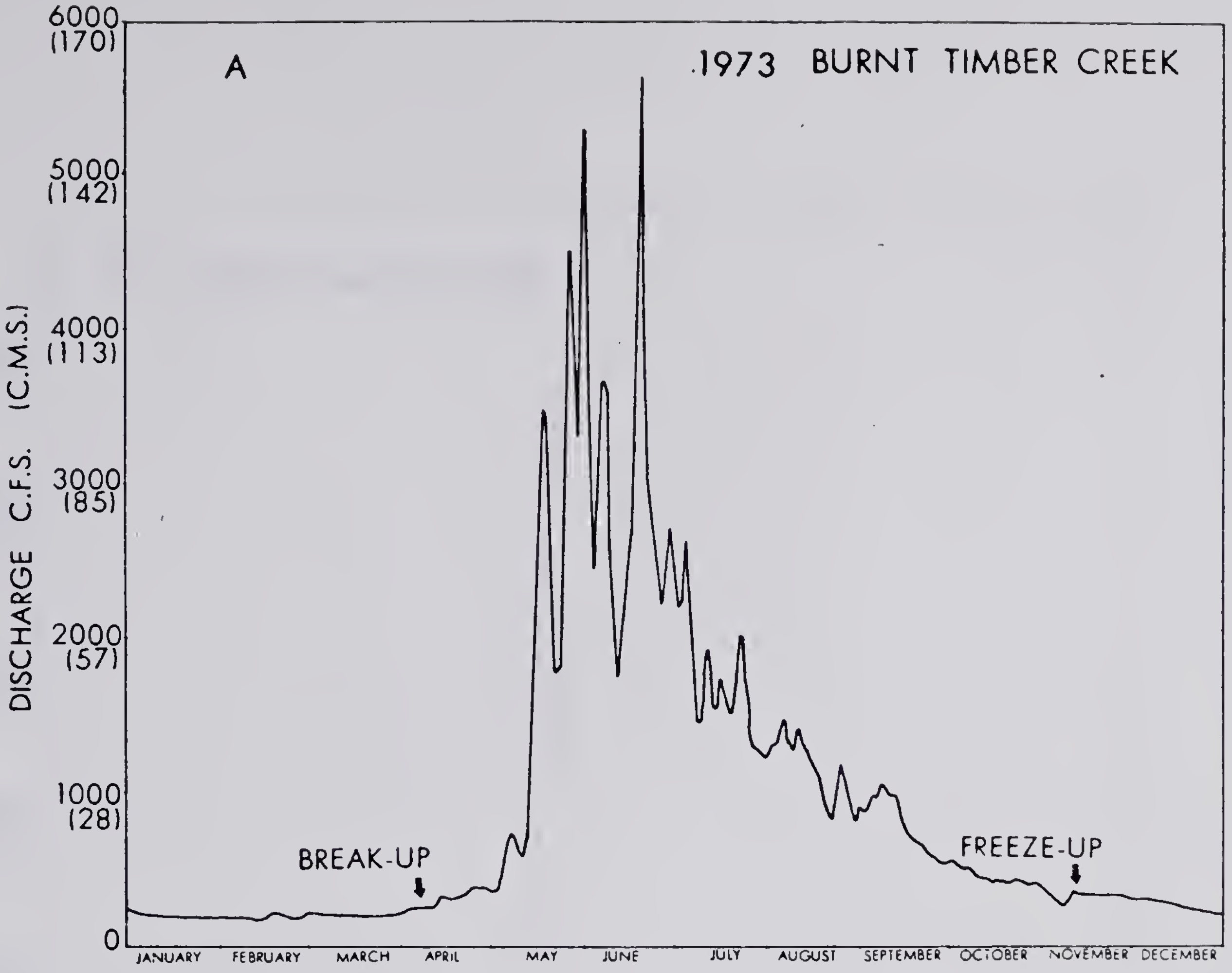
stations. Presently a manual gauge is in operation at Sundre, but this station is functional only during the ice free months, yielding incomplete data.

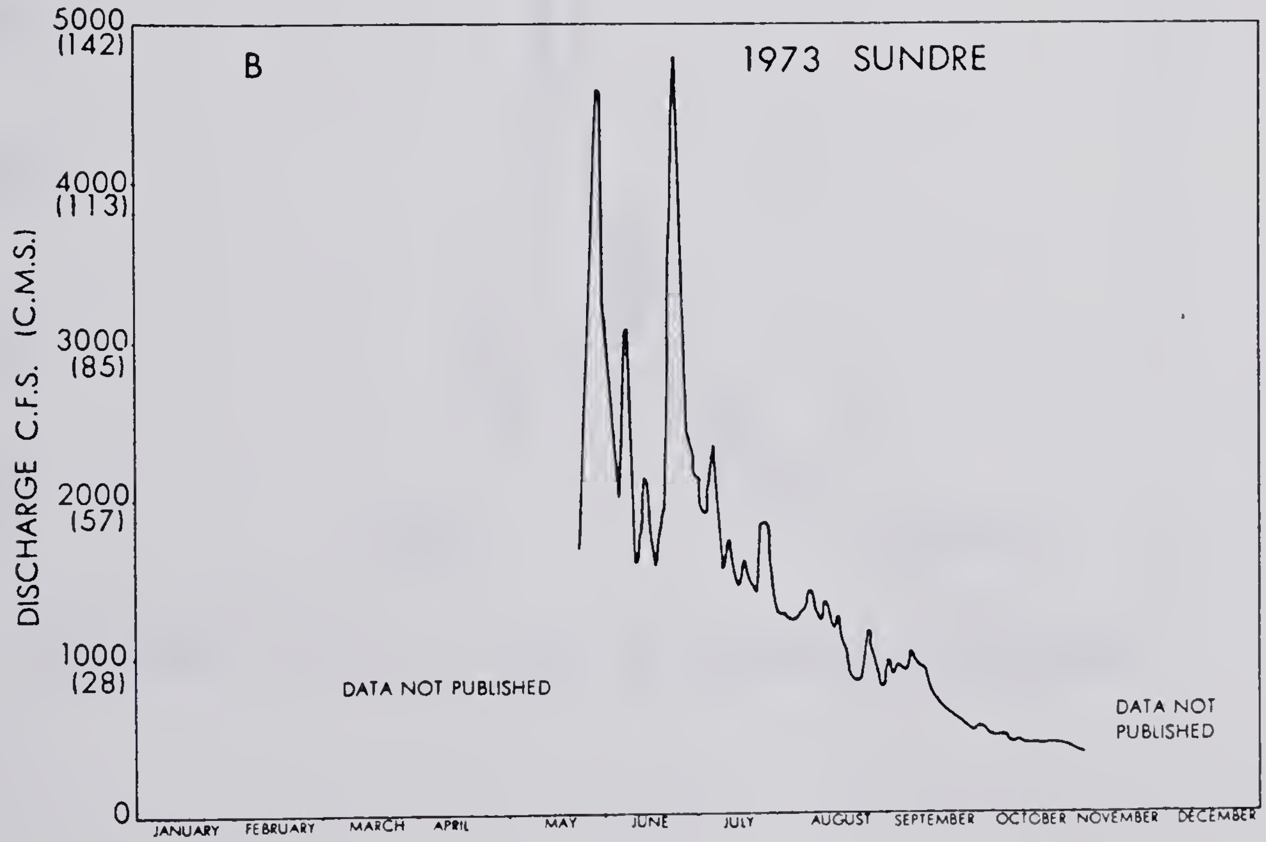
The hydrographs for 1973 at Burnt Timber and Sundre (Figures 2.1 A and B) showed similar discharge patterns. These included very rapid rises and a number of flood peaks during the months of May and June. After the final major peak there was a general decline in the discharge until freeze-up in the late fall and early winter. After freeze-up the discharge remained relatively constant until break-up in the spring. Superimposed on this general declining trend were smaller, less significant peaks. These were generally restricted to the months of July, August and early September.

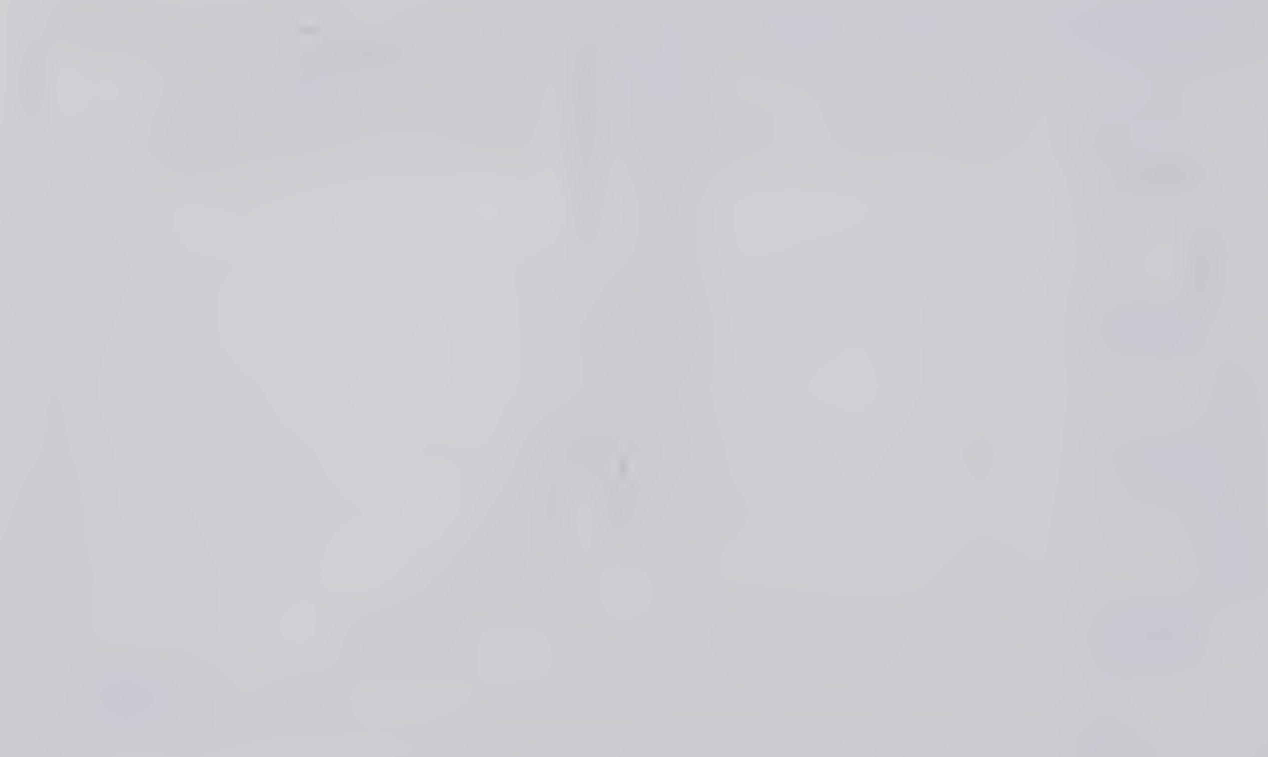
The magnitude, timing and number of the major flood waves vary from year to year. Figure 2.1 A. C. D. E and F show the Burnt Timber station hydrographs for the years 1973 to 1977. The same basic pattern, of large, short duration floods, is evident from these plots. The 1975 and 1976 hydrographs show that major floods occurred much later in the season and were generally of lower amplitude than in the other years. Possible reasons for this variation will be discussed below.

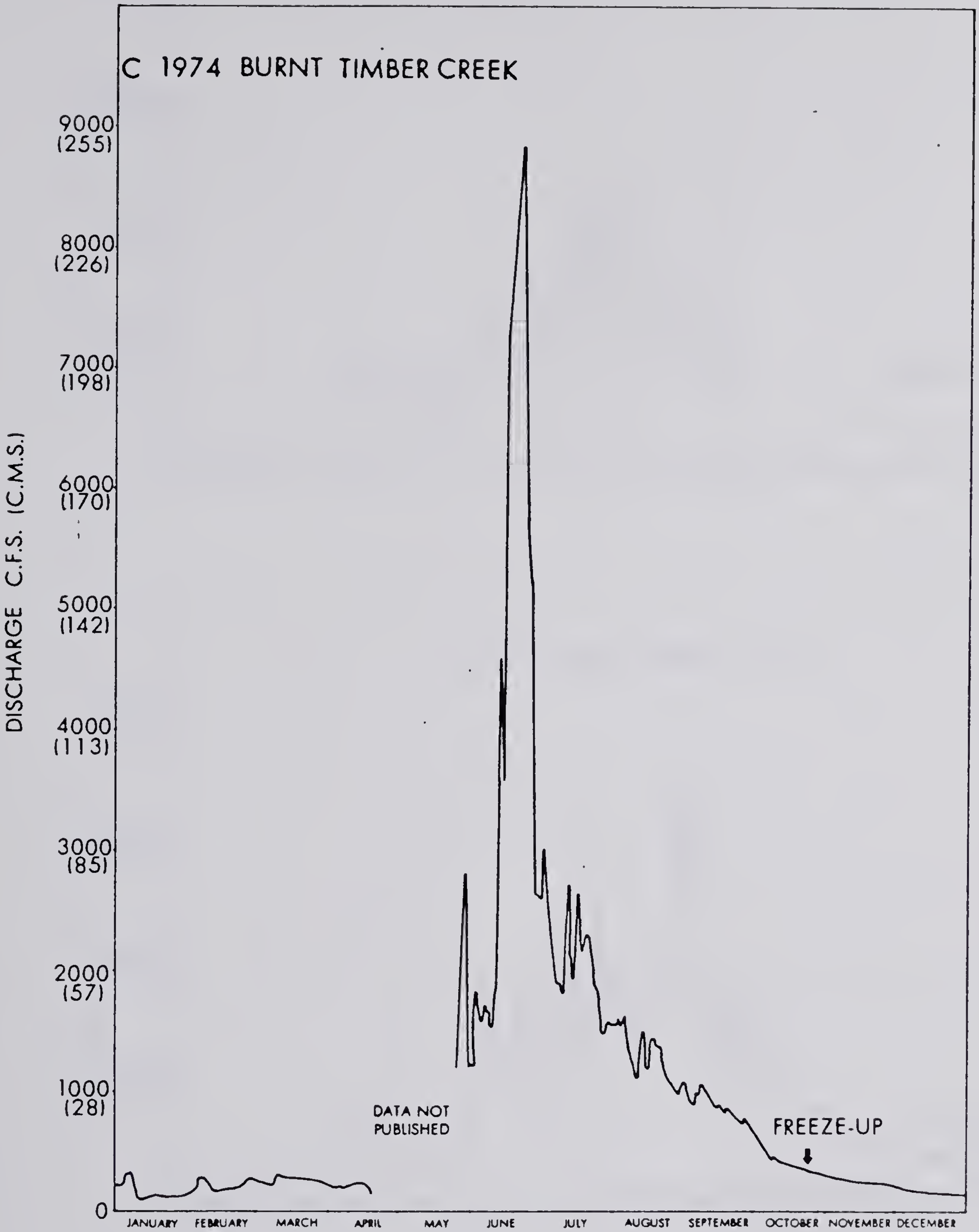
The hydrographs indicate a number of possible mechanisms for the generation of floodpeaks. The main peaks, occurring in early to midsummer (such as those recorded in 1973, 1974, and 1977), are caused by the addition of meltwater from the mountain snowpack. The magnitude and frequency of these peaks depend to a large degree on the climatic conditions prevalent at the time. Optimum conditions

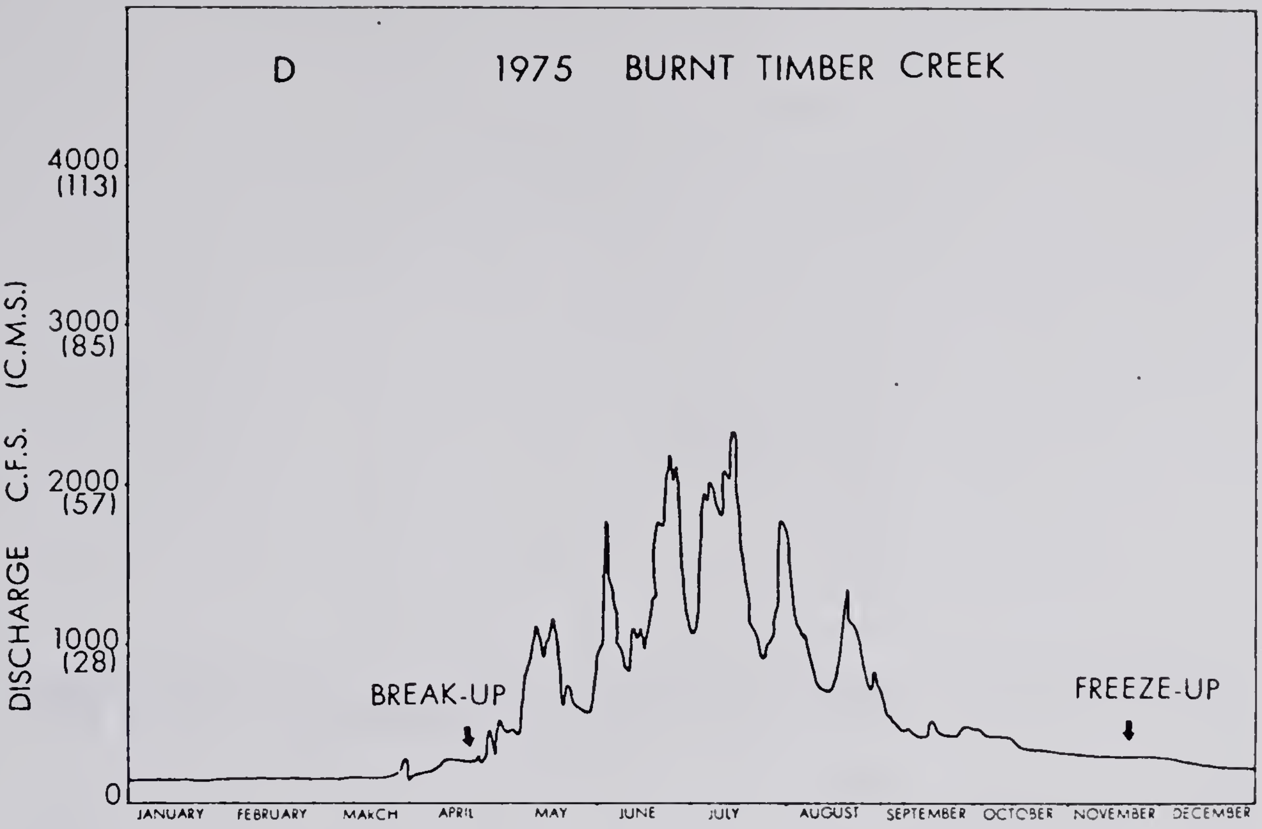
Figure 2.1 A to F. Discharge hydrographs based on mean daily discharges for the gauging stations at Burnt Timber Creek and Sundre for the years 1973 to 1977. (source: Canada, Department of Fisheries and Environment, 1973 to 1977)


A
1973 BURNT TIMBER CREEK
DISCHARGE C.F.S. (C.M.S.)
6000 (170)
5000 (142)
4000 (113)
3000 (85)
2000 (57)
1000 (28)
0
BREAK-UP
FREEZE-UP
JANUARY
FEBRUARY
MARCH
APRIL
MAY
JUNE
JULY
AUGUST
SEPTEMBER
OCTOBER
NOVEMBER
DECEMBER


B
1973 SUNDRE
DISCHARGE C.F.S. (C.M.S.)
5000 (142)
4000 (113)
3000 (85)
2000 (57)
1000 (28)
0
DATA NOT PUBLISHED
DATA NOT PUBLISHED
JANUARY
FEBRUARY
MARCH
APRIL
MAY
JUNE
JULY
AUGUST
SEPTEMBER
OCTOBER
NOVEMBER
DECEMBER




C 1974 BURNT TIMBER CREEK
DISCHARGE C.F.S. (C.M.S.)
9000 (255)
8000 (226)
7000 (198)
6000 (170)
5000 (142)
4000 (113)
3000 (85)
2000 (57)
1000 (28)
0
DATA NOT PUBLISHED
FREEZE-UP
JANUARY
FEBRUARY
MARCH
APRIL
MAY
JUNE
JULY
AUGUST
SEPTEMBER
OCTOBER
NOVEMBER
DECEMBER


D
1975 BURNT TIMBER CREEK
DISCHARGE C.F.S. (C.M.S.)
4000 (113)
3000 (85)
2000 (57)
1000 (28)
0
BREAK-UP
FREEZE-UP
JANUARY
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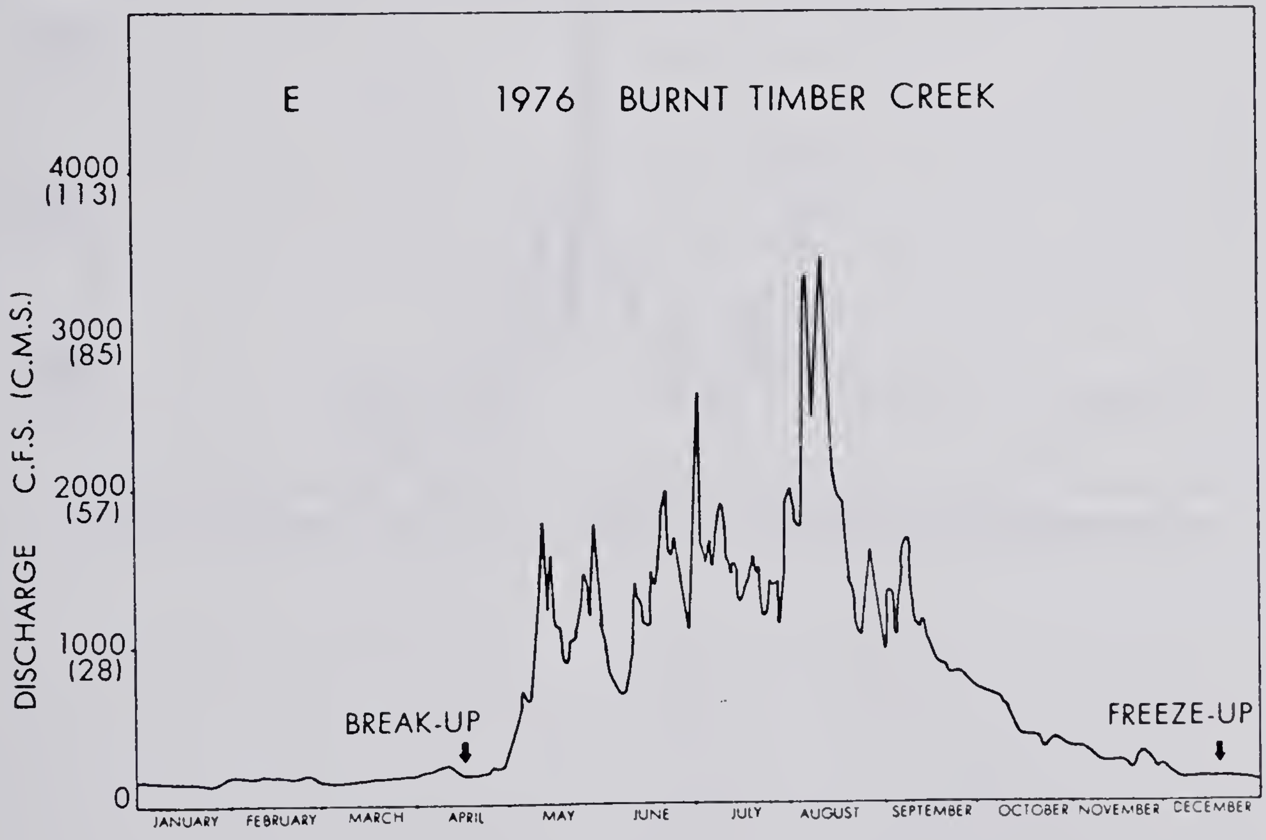
E
1976 BURNT TIMBER CREEK
DISCHARGE C.F.S. (C.M.S.)
4000 (113)
3000 (85)
2000 (57)
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BREAK-UP
FREEZE-UP
JANUARY
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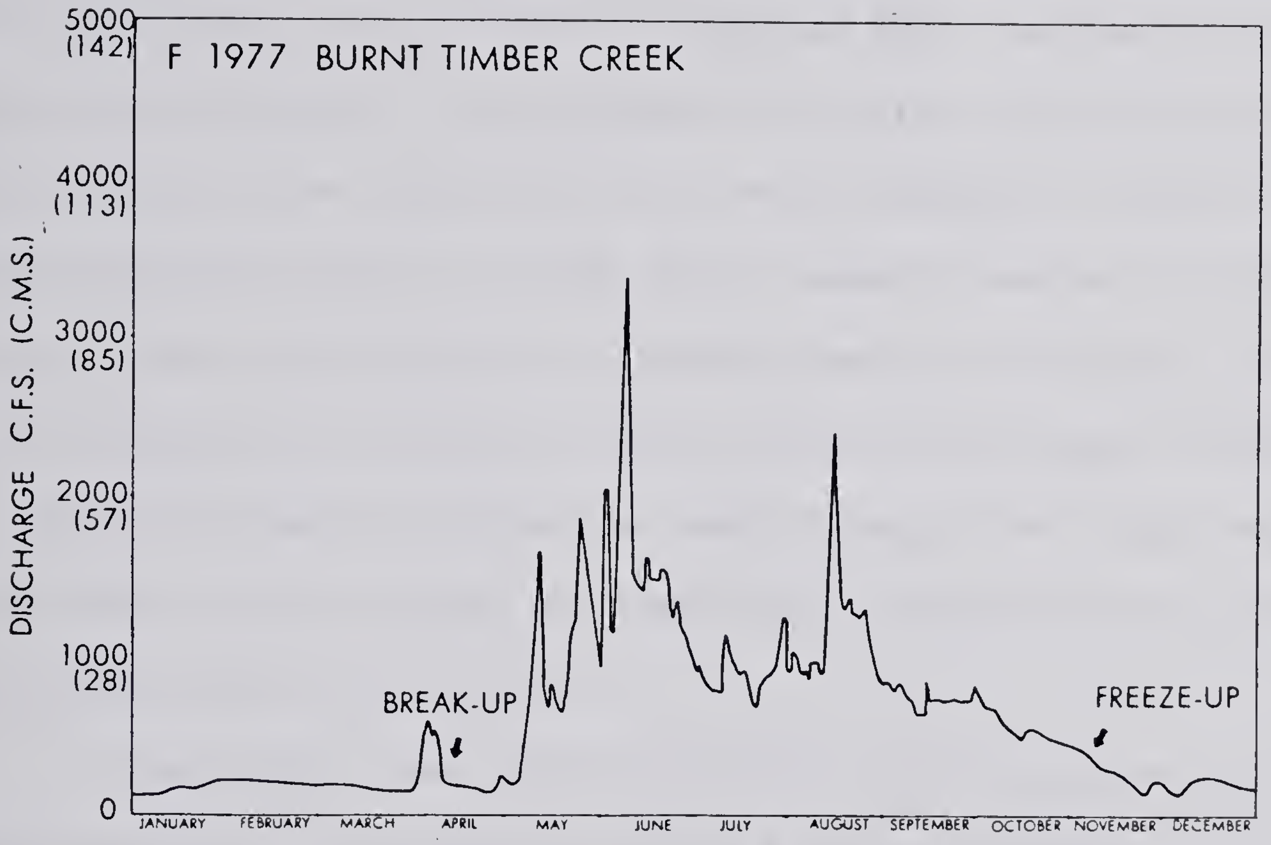
F 1977 BURNT TIMBER CREEK
DISCHARGE C.F.S. (C.M.S.)
5000 (142)
4000 (113)
3000 (85)
2000 (57)
1000 (28)
0
BREAK-UP
FREEZE-UP
JANUARY
FEBRUARY
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DECEMBER

include high temperatures combined with rain (Lowe and Sutton, 1977). These conditions will tend to accelerate the melting of the snow pack. Langbein (1940), described this type of flow pattern as "reservoir action", produced by the release of a floodwave from, for example, the rapid melting of snow and ice. Variations of the climatological conditions described above, in addition to the amount of snow contained within the pack at the time of the melt, will result in substantially different plots, such as those for 1976 and 1977. As can be seen from these two hydrographs, the flood waves occurring over a comparatively long period of time experience lower peak discharges. This may be explained by the gradual melting of the mountain snow pack. This will induce higher, total amounts of infiltration into the ground, leading to a more gradual streamflow release throughout the summer months. If the melt is rapid, infiltration capacities will be quickly exceeded, resulting in a large amount of runoff water entering the river over a relatively short period of time.

The smaller, less significant peaks occurring during mid to late summer are caused mainly by short duration convective storms, frequently experienced in this region. The related discharge peaks are usually of short duration and amplitude, resulting in less dramatic flood waves compared to those generated by snowmelt.

The gradual decline in discharge, after the passing of the main peak flows in the early summer, probably results from a gradual reduction in the rate of groundwater recharge. This recharge peaks shortly after the river peakflows occur, as can be seen in the 1973 hydrograph from Burnt Timber Creek. It is also interesting to note

that for the years where the main peak flows were subdued, such as in 1976, the seasonal falling limbs did not become markedly evident until late summer.

Very little has been written regarding the role of seepage of water from braided channels into the surrounding floodplain materials. This factor may have important implications with respect to the competence of a river from one reach to the next. Harrison and Clayton (1975) noted that substantial differences of competence may be experienced between a "gaining" and a "losing" reach. This may be of potential importance when considering the upper Red Deer River, especially where it flows from the foothills onto Bearberry Prairie. As will be discussed in the next chapter, Bearberry Prairie is a large gravel deposit bordering the Red Deer River for a distance of over twenty kilometres. These deposits are very receptive to water infiltrating from the active channels of the river. This fact may have significant implications for the dampening of the energy of the floodpeaks. The discharge peaks, as described above, rise rapidly during the early summer, at which point the ground water table would be comparatively low. A rapid rise of the river stage would, therefore, promote seepage into the surrounding Bearberry Prairie gravels. Recharge of some of the water back into the Red Deer River, either directly or via Bearberry Creek, would occur, but presumably later after the passing of the flood wave. The hydrographs for 1973, at Burnt Timber Creek and Sundre, indicate that this phenomenon may be significant. For individual flow events there are consistent decreases in the discharge at Sundre relative to the Burnt Timber station.

The flows during the second half of July, to the end of the period of record at Sundre, equal those at Burnt Timber Creek. This suggests that a mid to late summer equilibrium between groundwater and river discharges is attained. The early summer discrepancy of discharges measured for the two stations may be more apparent than real, however, because of the instability of the channel at Sundre. This may cause inaccurate discharge readings for that station. A detailed examination of ground-water flow patterns and fluctuations would be necessary before realistic conclusions regarding this problem could be formulated.

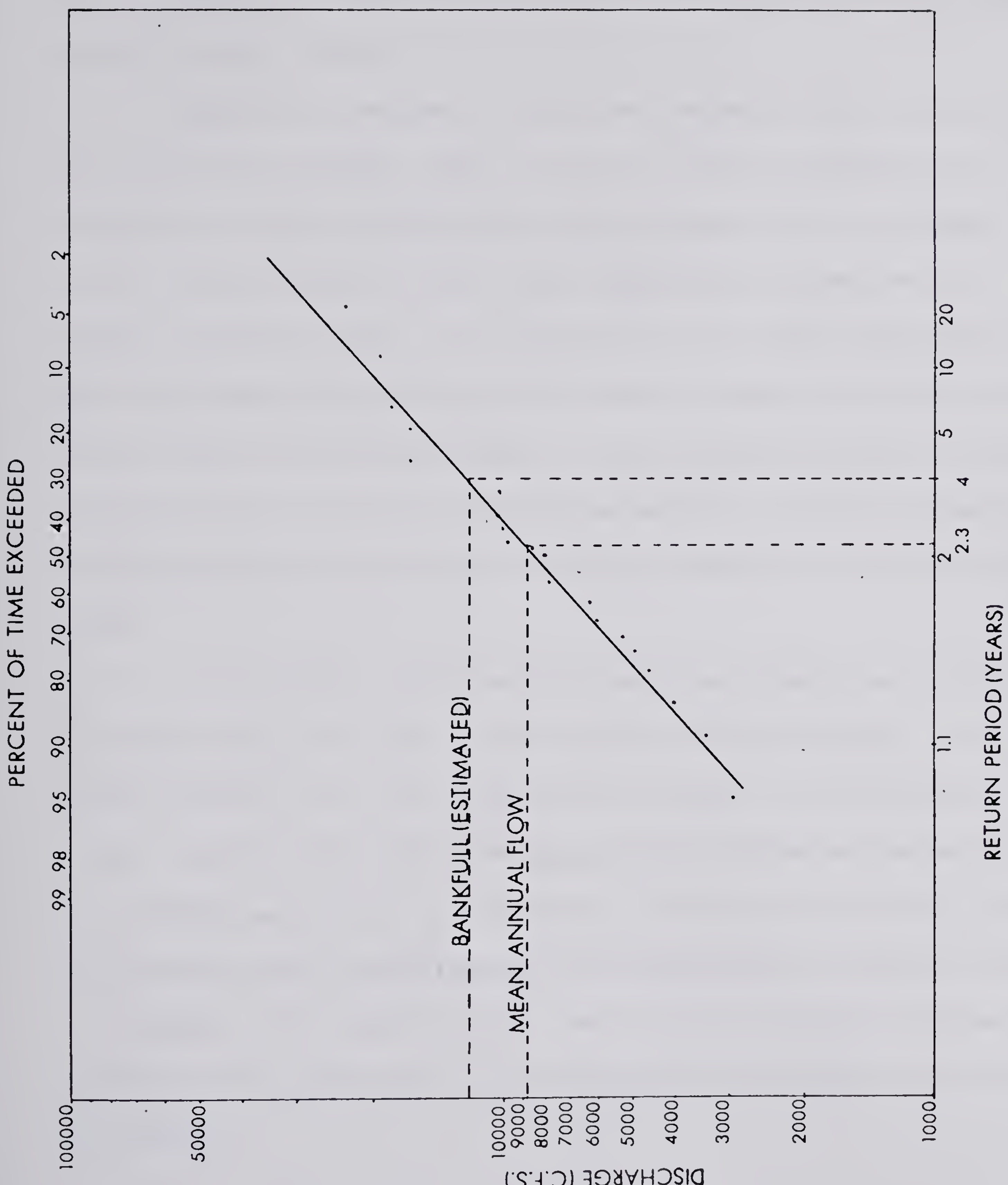
A discharge duration curve (Figure 2.2) was constructed for the data obtained from the Sundre station. The number of data years is small, so that a reasonably accurate examination may be made only for the short return period flows. The method used to obtain this curve was outlined by Dalrymple (1960) and is defined by:

$$T = \frac{N + 1}{M} \qquad (1)$$

where T is the return period in years, N is the length of the record, and M is the rank of an individual flow event. Dalrymple (1960, p. 16) stated that;

> Recurrence levels computed by this formula give results similar to those computed by the California method, but lack the theoretical deficiencies of the latter. (It) is simple to complete, is applicable to annual flood data and the partial duration series.

Figure 2.2 Discharge duration curve for the gauging station at Sundre. (source: Canada Department of Fisheries and Environment, 1977)



1 C.F.S. = 0.028 C.M.S.

This method, incorporating the maximum annual instantaneous discharges, was used instead of the partial duration method because of the similarity of results to those from records of more than ten years in length (Leopold et. al., 1964).

Bankfull discharges at Sundre, estimated by Alberta Environment (Lowe, pers. comm., 1979), are in the order of 12,000 c.f.s. (340 c.m.s.), which corresponds to the four year flood (see Figure 2.2). This is within the time limit suggested by Leopold et. al., (1964) and other workers. The high frequency of this stage implies that the channel surfaces within the bankfull range do not have sufficient time to become stabilised by vegetal growth and that the bed material will be exposed and highly susceptible to fluvial processes inducing erosion and deposition during the periods of high flow (Harvey, 1969).

In conclusion, the upper Red Deer River experiences frequent high discharge events and a significantly variable discharge. The peaks, however, also vary in magnitude from year to year depending on the climatic conditions leading up to and influencing the melt of the mountain snow pack. As indicated in Chapter One, this is of potential significance in contributing to the development of braided channel patterns. The relative effect and importance of the discharge variable will be discussed in greater detail in the final chapter of this thesis.

2.3 CLIMATE AND VEGETATION

The climate of the study area is classified as continental, with warm summers and cold winters (Canada Land Inventory, 1972). Precipitation varies in quantity from 45 cm in the east at Sundre, to over 50 cm in the west. A majority falls during the summer months, although a significant amount falls as snow in the foothills and mountain regions (Boydell, 1970). A number of convectional storms occur in the summer, during the time from mid-July to mid-August. These storms are of a localised nature and may result in destructive hailstorms (Department of Transportation, 1967). The months of June and July experience the greatest average amounts of precipitation.

The flora in the study area have been grouped into the class of Boreal Sub-alpine Transition Forest by North (1976). The dominant species found include Engelmann spruce (*Picea engelmanni*) with lodgepole pine (*Pinus contorta*) in areas which have had a recent fire history (Canada Land Inventory, 1974). North (1976) also included white and black spruce (*Picea glauca*, *Picea mariana*) and aspen poplar (*Populus tremuloides*) as important species to be found in this ecotone between the boreal and the sub-alpine forests.

In the river valleys, with relatively damp soils, white spruce dominates. These sites may have spruce trees ranging in age up to 240 years (North, 1976). The areas which are covered with till are dominated by aspen poplar and spruce species. The areas of higher elevation are dominated by lodgepole pine.

Because of the high soil moisture content, the spruce and

poplar dominate in the areas bordering streams. The high moisture content, and the effect of the rivers acting as barriers, inhibits fire from destroying the vegetation in these areas. In less sheltered locations, such as the active floodplains or on exposed gravel surfaces, willows and alders (spp. salix and spp. alnus) are dominant (North, 1976). A tentative succession of willows and alders, followed by the poplars and spruce species, is therefore expected for these areas. As North (1976) states, "...the evidence of present forest succession clearly points to the climax position of white spruce in mesic sites." (North, 1976, p. 35).

CHAPTER THREE

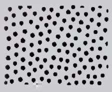
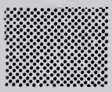
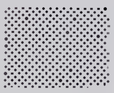
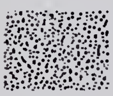
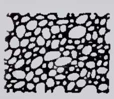
BEDROCK AND SURFICIAL GEOLOGY

3.1 INTRODUCTION

The surficial deposits and the general geomorphology of the study area, immediately surrounding the study reach, were assessed to satisfy two main objectives. First, it was necessary to map the surficial deposits of the area to identify the possible sources of coarse debris for the upper Red Deer River. Second, it was necessary to interpret the effects of the geomorphic history on the morphology of the present day river floodplain. An attempt was made, therefore, to refine information gathered in the field, supplemented by the previous studies of McPherson (1963) and Boydell (1970, 1978), to establish the probable late glacial and post glacial history of the study area.

A preliminary interpretation of aerial photographs, prior to the field work, resulted in the delineation of major geomorphic units based on the morphological characteristics of their surfaces. These units were then field checked from exposures for homogeneity in terms of constituent materials. A final interpretation of the aerial photographs was then carried out. The derived information was mapped directly onto the photographs and later transferred to a topographic map of scale 1:225,000 (Figure 3.1).

LEGEND

UNIT	SUBUNIT	MORPHOLOGICAL DESCRIPTION	SURFICIAL MATERIAL	UNDERLYING MATERIAL
	1	Multiple channel flood-plain	Gravel	Varied
	2	Single channel flood-plain	Gravel	Varied
	3	Single channel flood-plain	Sand, silt, clay	Varied
	1	Flat terrace tread	Alluvial sand, silt, clay	Gravel
	2	Flat terrace tread	Alluvial sand, silt, clay	Till
	3	Convex terrace tread	Alluvial gravel	Varied
	1	Hummocky valley fill	Eolian silt and sand	Till
	2	Hummocky valley fill	Till	Varied
	3	Gently undulating valley fill	Lacustrine silts and clays	Till
	1	Abandoned channel	Gravel, sand, silt, and clay	Bedrock
	2	Abandoned channel	Gravel, sand, silt, and clay	Various unconsolidated materials
	1	Discontinuous till veneer	Till	Bedrock
	2	Hummocky moraine	Till	Varied
		Lacustrine plain	Highly variable-till, sand, silt, clay	Varied
		Undulating outwash fan or delta, some poorly defined abandoned channels.	Eolian sands- cover sands with some modified sand dunes. Organics in depressions.	Varied
	1	Colluvial slope, inactive deepseated massmovements	Mainly unconsolidated material. Some mudflows occurring in bedrock.	Varied
	2	Colluvial slope, active superficial sheetwash and mudflows		Varied
+++++ +++++ +++++ +++++		Erosional remnant	Varied-eolian sands and till	Till, bedrock

—— .... Geomorphic Boundary; Defined, Approximate

——3500—— Contour (Feet) (1 Foot=0.305 Metres)

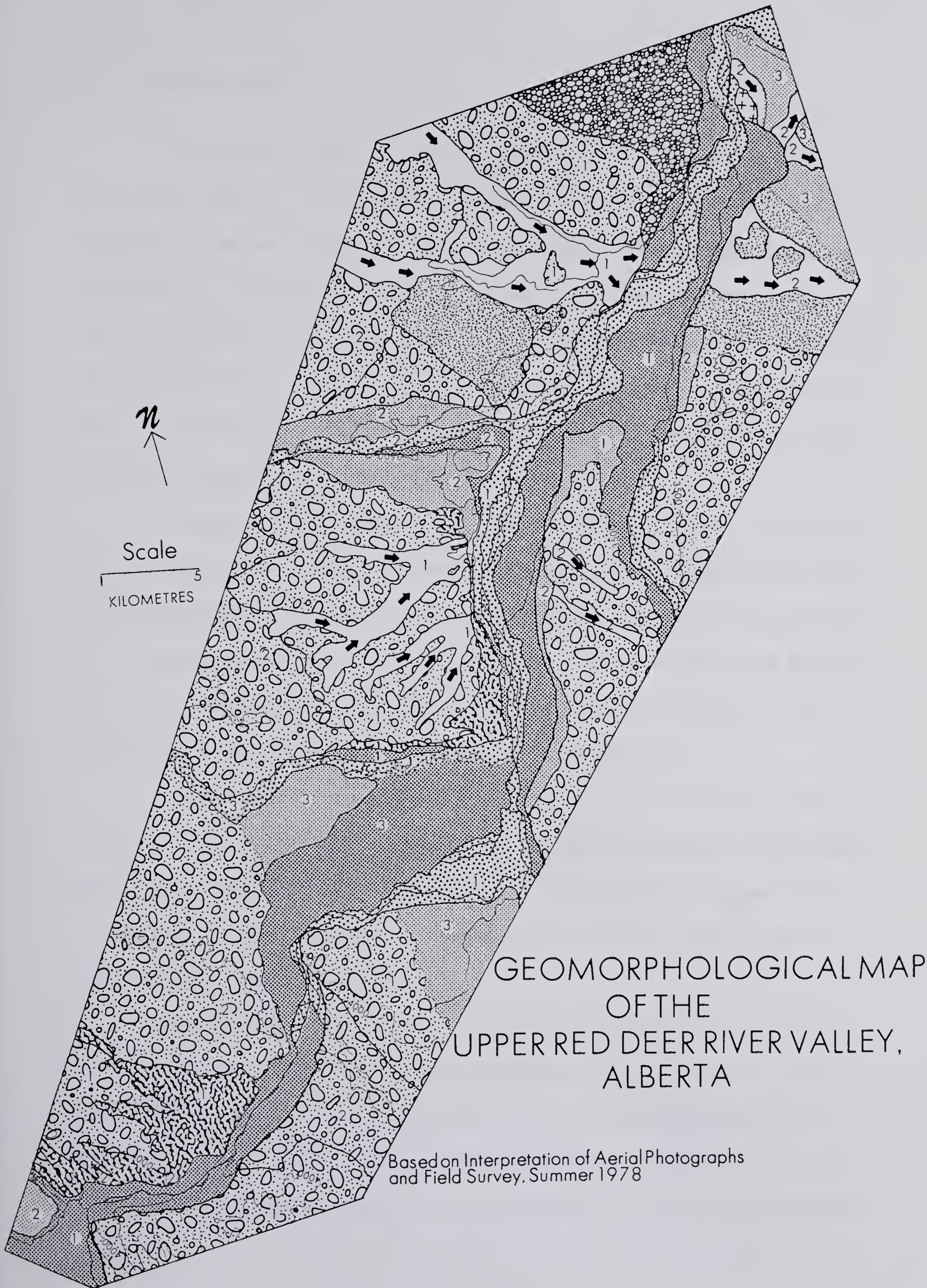

Scale
1 5
KILOMETRES
GEOMORPHOLOGICAL MAP
OF THE
UPPER RED DEER RIVER VALLEY,
ALBERTA
Based on Interpretation of Aerial Photographs
and Field Survey, Summer 1978

Figure 3.1

3.2 BEDROCK GEOLOGY

The bedrock geology was not checked in the field because of time constraints. The previous accounts of Mackay (1937) and Beach (1940) are the bases of the following outline. The bedrock geology comprises four formations ranging in age from Upper Cretaceous to Tertiary (Figure 3.2). The oldest is the Wapiabi Formation, characterised by greenish, sandy shales with numerous interbeds of sandstones and clay ironstones. Exposures of this formation are limited to the William's Creek area in the west.

The Belly River Formation, also of Cretaceous age, is composed of white, grey and brown sandstones interbedded with light green and black shales and occasional lenses of conglomerates containing quartzitic and chert pebbles. Some thin beds of clay ironstone and bentonite are also included. This formation is well represented in the field area from William's Creek east to Coalcamp Creek.

The youngest Cretaceous deposit is that of the Edmonton Formation. This is composed of light buff to grey sandstones interbedded with dark brown sandy and blocky shales as well as a few minor beds of bentonite. It is limited in extent and occupies a zone along the eastern slopes of the foothills.

The youngest major unit is the Paskapoo Formation, which is of Tertiary age. This is located to the east of the foothills and underlies the remainder of the study area. It is characterised by hard, light grey, buff and yellow-brown sandstones with light bluish-grey, dark olive-green, sandy and blocky shales. A few beds of bento-

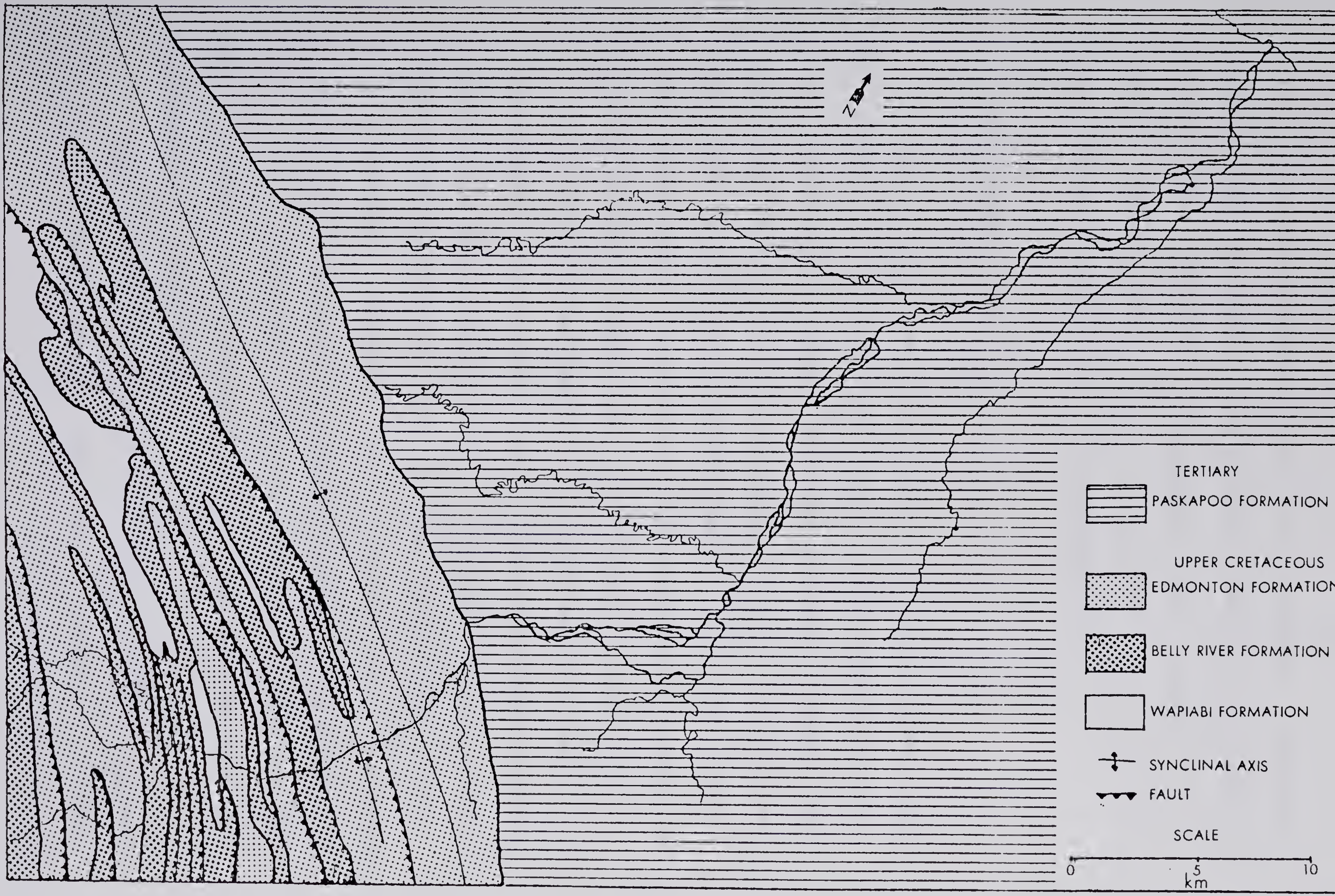


Figure 3.2 Bedrock geology of the study area.

nitic and carbonaceous shales occur in this formation. Exposures are confined mainly to the upper Red Deer River Valley but a number of highly weathered exposures on topographic highs were noted in some roadcuts. The weathering products of this material in many cases are almost indistinguishable from the overlying till.

3.3 PHYSIOGRAPHIC DIVISIONS

The study area may be divided into a number of physiographic units (Figure 3.3). These include structurally controlled ridges in the foothills, for example the Stafne and Parker Ridges, which strike in northwest-southeast directions and are located between the western edge of the Bearberry Prairie and William's Creek. These ridges run obliquely to the Red Deer River and are small anticlinal structures composed of Belly River and Edmonton Formation rocks.

The second physiographic unit is the Bearberry Prairie. This is an extensive zone of gravel contained within a markedly widened portion of the Red Deer River Valley. It is expressed by surfaces at two different elevations, with the lower surface (B.B.1) being the most extensive of the two. The upper surface (B.B.2) is located in the northwest portion of the Bearberry Prairie. The southern and eastern limits of Bearberry Prairie are marked by the Red Deer River while the west is bounded by Stafne Ridge. The northern limit is demarcated by Bearberry Creek.

The third unit is composed of a highlands area to the north and east of Bearberry Prairie. This unit is dissected by a number

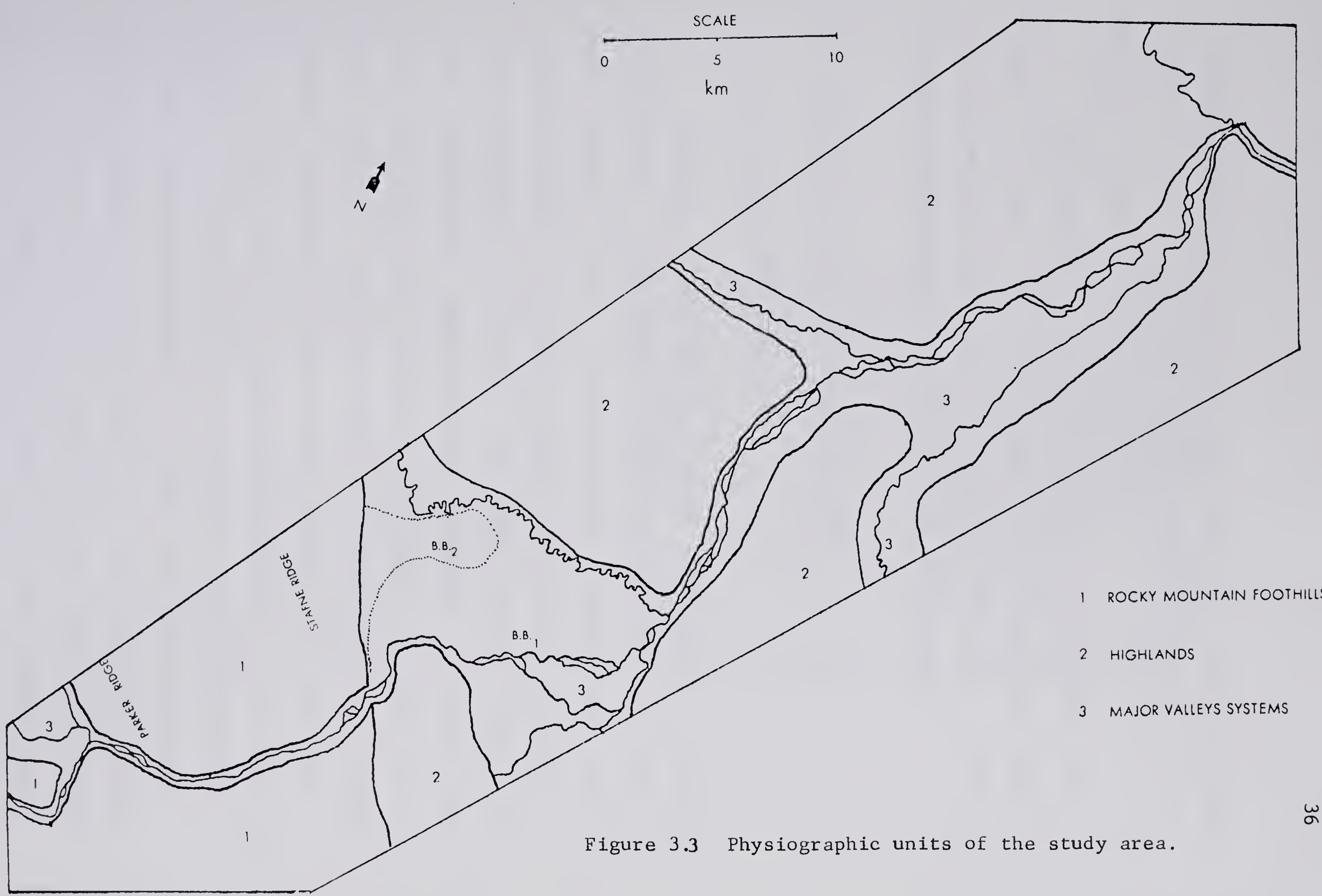


Figure 3.3 Physiographic units of the study area.

of streams including James River, Eagle Creek and a number of unnamed creeks, as well as several dry channels, some of which have been, in the past, deeply incised into the surface.

The remaining physiographic division is the Red Deer River Valley. This unit extends from southwest to northeast and transects the units mentioned above. It is of variable width and in places is entrenched approximately 60 metres into the local bedrock.

3.4 SURFICIAL DEPOSITS AND GEOMORPHIC UNITS

3.4.1 LOWER CORDILLERAN TILL

The oldest till found in the study area has quite distinctive characteristics. Perhaps most important is the consistently high gravel component of the till, where observed. The gravels are well rounded and appear to be wholly composed of Cordilleran materials, namely limestones and quartzites. Some minor striations were found on the larger clasts, especially on the limestones. The matrix has a high clay content with secondary silt and sand. The deposit is highly indurated or compacted (Figure 3.4). Three exposures of this material were found and these are shown as sites 1 to 3 in Figure 3.5. At site 1, the till directly overlies bedrock. In the other two locations the depth to bedrock is not known. The contacts between this till and overlying deposits are sharp, indicating that little post depositional disturbance has occurred. The age of this material is not known but it may be a correlative of the Albertan till reported in southwestern Alberta by Alley (1973) and Stalker and Harrison

Figure 3.4 Stratigraphic section in Red Deer River Valley showing unconsolidated materials. The highest and lowest units are Cordilleran tills separated by lacustrine sediments.

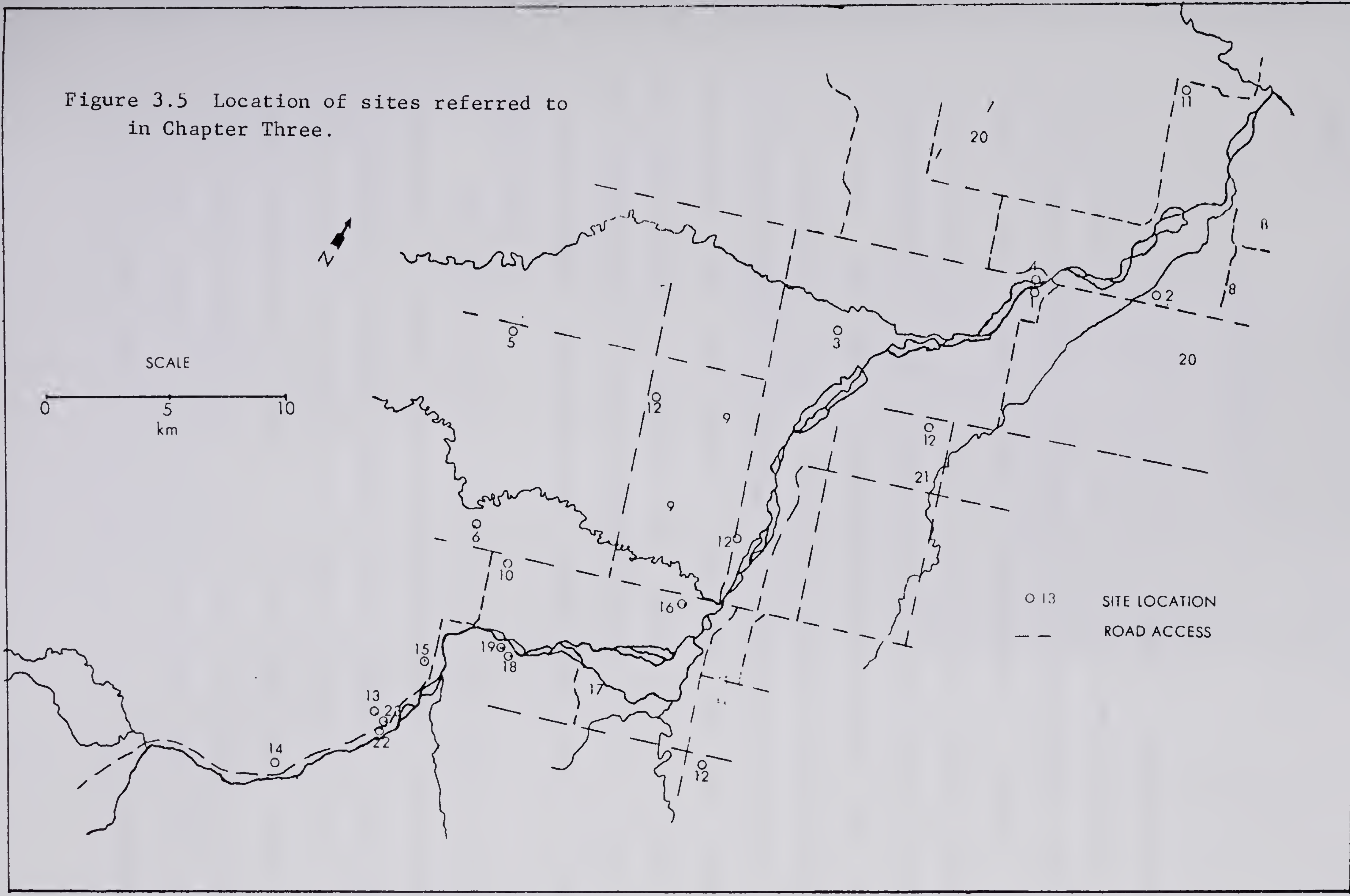


Figure 3.5 Location of sites referred to in Chapter Three.

(1977).

3.4.2 INTERTILL DEPOSITS

Intertill deposits were clearly preserved in two exposures, and possibly in a third. The first two (sites 1 and 3 in Figure 3.5) consist of rhythmically deposited lacustrine sediments separating two tills. These rhythmites are approximately 1.2 metres in total thickness, although this varies considerably along the length of the exposures. They are composed of alternating beds of sand and silty clay. The thickness of the sand beds is variable, with a maximum of 15 centimetres. These coarser grained beds show slightly variable sorting characteristics with no apparent sedimentary structures. Some larger clasts are imbedded in these units. The silty clay beds are highly variable in thickness with a maximum of 20 centimetres. The sorting appeared to be good throughout with a few occurrences of small pebbles. Very fine parallel laminations occurred in these units (Figure 3.6).

The exposure at site 1 (Figure 3.5) is the largest, having been exposed by lateral erosion of the Red Deer River. In a horizontal direction the intertill deposits are not extensive. They become thinner and gradually disappear over a distance of 20 to 25 metres after which the two tills are in contact with each other. At site 3 only a small exposure exists, so the extent of the intertill material is not known. However, these relationships suggest that ponding of a localized nature existed. The beds of coarse sand and fine gravel

Figure 3.6 Close-up of the lacustrine sediments seen in Figure 3.4. Note the different textural characteristics of the separate beds.

show that a relatively high energy sedimentary environment existed at the time of deposition. These beds are in contact with the silty clays above and below, with no transitional phases or upward fining sequences noted.

The third section contained gravels overlain by till (site 4 in Figure 3.5). The section description is included here because it is not known whether there was till deposited over bedrock prior to the deposition of these gravels. Presently the gravels overlie bedrock but fluvial erosion of a possible former till may have occurred prior to alluvial deposition of the gravels. The gravels are moderately sorted and show some bedding. Small lenses of well sorted sands and fine gravels are incorporated into the deposit. There is a weak preferred orientation of clast long axes towards the east-southeast. The lithology of the material suggests a Cordilleran origin, as the majority of the gravels are quartzites with no apparent Shield rocks. The matrix consists of medium to fine sands with an apparently insignificant silt and clay content (Figure 3.7 A and B).

3.4.3 UPPER CORDILLERAN TILL

The upper unit of the two Cordilleran tills identified occurs extensively within the western and central portions of the study area. It constitutes two main geomorphic types; hummocky moraine

Figure 3.7 A Exposure along the Red Deer River containing gravels. These gravels (1) are located stratigraphically below the upper Cordilleran till.

Figure 3.7 B Close-up of the exposure shown in Figure 3.7 A. Note the small lenses of fine gravel interbedded in the coarser clasts.

merging into the foothills region to the west, and a discontinuous, thin, till veneer over bedrock in the central parts of the area. This till is fairly heterogeneous in its constituency, ranging from reworked, fine lacustrine sediments with occasional pebbles, to stony tills with a high content of quartzite and limestone, Cordilleran gravels. The subrounded to rounded shapes of these clasts indicate that this fraction had been fluvially transported prior to till incorporation. A small exposure in the hummocky moraine (site 5 in Figure 3.5) showed a very fine grained matrix with small amounts of gravel. Cordilleran gravels were found there as were a large number of local bedrock fragments. The local bedrock clasts are very weak, easily broken by hand and their proportionately large amount suggests that transport was not over extensive distances prior to deposition. In the hummocky moraine on the western edge of Bearberry Prairie (B.B.2) there is evidence of localized ponding and topographic reversals (Figure 3.8). Several layers of silty clay, light brown in colour, and up to 15 centimetres thick, alternating with dark grey clay lenses approximately two to three centimetres thick were found. The beds are highly distorted and merge with till at the edges of the hummocks. Occasional dropstones occur in the rhythmite deposits. The topography of the hummocky moraine is highly undulating with the local relief of the hummocks increasing in a westerly direction. Figure 3.9 shows a section extending to bedrock in this area (site 6, Figure 3.5).

The second stratigraphic unit, composed of the upper till, is that of the discontinuous veneer over bedrock. In most areas

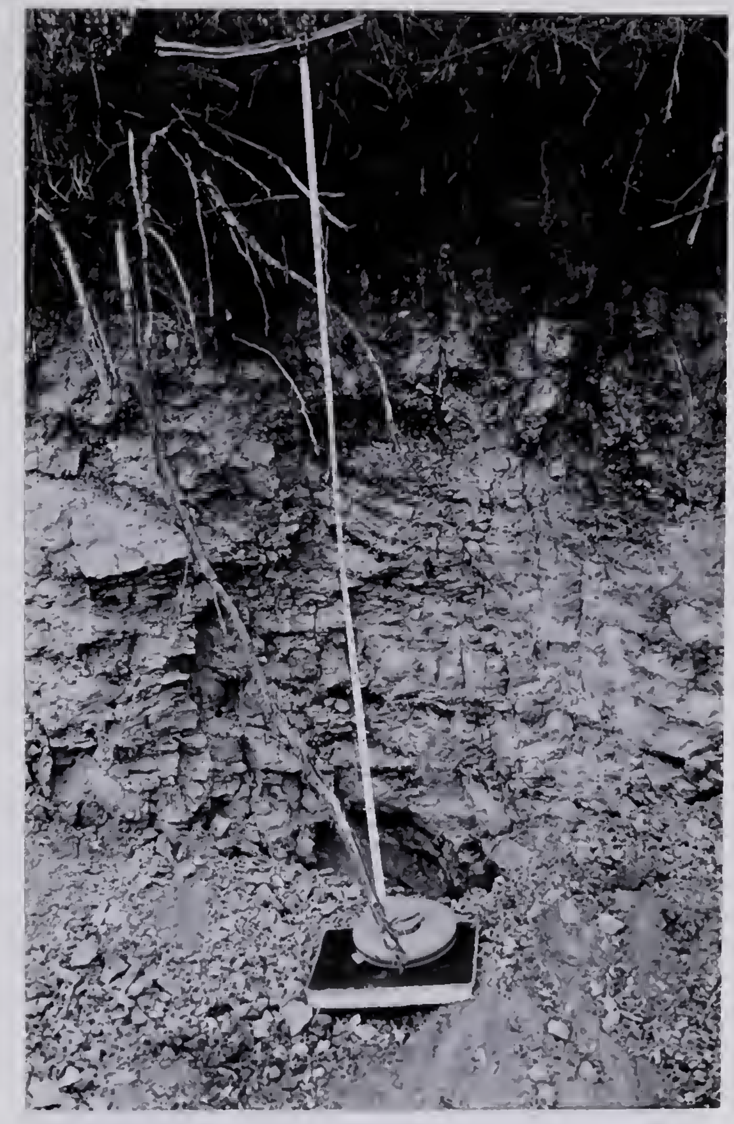


Figure 3.8 Ponding sediments found in a topographic high (knoll) in the hummocky moraine.

Figure 3.9 Exposure in the hummocky moraine extending to bedrock. Difference in tone in the upper portions may be result of pedologic processes.

the surface morphology of this unit is gently undulating with the relative relief of the knolls being considerably less than that found in the hummocky moraine. Figures 3.10 A and B and 3.11 illustrate these comparative morphologies. The till cover is generally thin with bedrock occurring in a number of road cuts. The nature of this material is consistent with that found in the hummocky moraine. Gravel contents vary over short distances and the till has a very dominant clay matrix. Angular fragments of the local bedrock are quite numerous.

The differences noted between the upper and lower tills are mainly in the degree of compaction and the clast content. As previously mentioned, the lower till is highly compacted or indurated whereas the upper till has a much looser packing. The upper till has a fine blocky structure while the structure of the lower till, although blocky, is much coarser. The lithologies of the two tills are consistent in that they are both Cordilleran in origin. However, the upper till contains appreciable amounts of local bedrock, a feature that is largely missing from the lower unit. A final comparative point is that of colour. The lower till is a grey-brown to grey colour, while the upper one is predominantly light to dark brown. The lower till appears to have undergone a certain amount of chemical reduction, while oxidation appears to have been dominant in the upper unit.

Several clasts of Canadian Shield granites were found in some till exposures. The visual characteristics of the tills, where these indicator materials were found, did not differ significantly from

Figure 3.10 A and B Oblique, aerial photographs of the hummocky moraine.

A

B

Figure 3.11 Oblique, aerial photograph of the till veneer landscape.

Figure 3.12 Erratic found in the study area.

the Cordilleran tills previously discussed, thus a differentiation based on superficial characteristics is not possible. Nowhere was there a marked concentration of Canadian Shield rocks but, rather, they occurred singly or in relatively minor proportions when compared with the Cordilleran materials. It is possible that Shield stones were transported into the area, from initial, Laurentide, depositional zones northwest of the study area. Such transport could have been caused by ice of the so-called Cordilleran "Athabasca Lobe", or the "Saskatchewan Lobe" as Boydell (1978) referred to it. This ice was also largely responsible for the creation of the Foothills Erratics Train (Stalker, 1956). A number of boulders belonging to this Foothills Erratics Train were found on the surface (Figure 3.12), mainly to the east of the Red Deer River near Sundre. Erratics further west have been noted, however, by Stalker (1956), McPherson (1963), and Boydell (1970).

3.4.4 SPILLWAY CHANNELS

Several large channels, eroded into bedrock, are concentrated in the northwestern section of the study area, north of James River. Their morphological characteristics include broad flat-bottomed, steep sided, linear depressions aligned in a southeast down-valley direction. The depth of these features varies from 30 to 50 metres below the highlands surface. Their depths decrease towards the southeast and the channels become less distinct as they merge into the Red Deer River Valley. The materials found in these channels vary from

gravel-rich masses to fines, containing sand, silt and clay. The channel bottoms in most cases are covered with muskeg. The materials underlying the muskeg accumulations are not known but in some other locations the bottom morphology is gently undulating and composed of gravels in a sandy matrix. The locations of these channels are shown in Figure 3.5, (site 7). Figure 3.13 A and B are oblique aerial photographs of two of the channels.

In two of the channels three distinct terraces were noted at different elevations. It was not possible to determine the existence of any sedimentary sequences or to differentiate between materials contained within them, because of the lack of exposures and heavy vegetative cover. The possible significance of these terraces will be discussed in a later section.

Abandoned channels of a second type are located on the eastern side of the Red Deer River, to the east of Garrington Bridge (site 8 in Figure 3.5). They are broad, flat, linear depressions and in some cases are very shallow and indistinct. They are clearly evident, though, on the aerial photographs. These channels appear to have been incised into unconsolidated materials and occur at varying heights above the river. A number of minor topographic highs are contained within the channels with local relief of approximately 5 metres between the bottom of the channels and the top of the highs. Some gravel is evident on the channel bottoms but is conspicuously absent on the elevated surfaces. The trend of these features is from northwest to southeast, with a very gentle slope toward the southeast. The better developed lower channels of this set are located approximately 8 to

Figure 3.13 A and B Meltwater channels incised into the highlands surface.

A

B

10 metres above the present Red Deer River floodplain level. They are 2 to 3 metres deep with well developed channel forms. The channel bottoms contain muskeg making it impossible to determine the nature of the underlying material. However, a high gravel content was noted in a small road cut in one of these depressions. The orientation of these channels varies considerably but all tend to slope approximately eastward, away from the Red Deer River.

A third type of abandoned channel occurs on the west side of the Red Deer River between Sundre and the James River, incised into the highland surface (site 9, Figure 3.5). The channels extend from the margin of the hummocky moraine in the west and slope down to the Red Deer River. The channels are up to one kilometre wide and attain a maximum depth of approximately 30 metres near their mouths. They have the appearance of gully systems with short, branched headwater sectors, and are approximately U-shaped in appearance. This shape is not very distinct because colluvial processes along the walls have produced gently undulating sides. At present, small underfit streams occupy the valley bottoms. Some of the materials into which these channels are incised are similar to the tills found on the interfluves. These materials are also found along the valley sides and are presumably tills relocated by mass movement processes. The valley bottoms also contain fine grained alluvium.

3.4.5 OUTWASH AND ICE-CONTACT DEPOSITS

Two major relict fan deposits, composed mainly of gravels and sands, have been identified in the field area. The first comprises the B.B.2 surface. Only a small remnant of the original surface remains undisturbed. At site 10, Figure 3.5, the gravels underlying this surface are moderately well sorted with a sand matrix. There are, abrupt vertical-plane variations in the sorting characteristics of the material, with lenses of fine sand occurring within the gravel beds. The depth of the exposed gravels is approximately 5 metres under which there occurs a bed of silt and clay. The depth of this bed and the nature of the underlying materials are not known. The preferred orientation of the gravel indicates that the dominant paleoflow direction was from the south, that is, from the present Red Deer River Valley, as it emerges from the foothills. This gravel deposit occupies only a very small portion of the B.B.2 surface. The remainder is composed of lacustrine fines and till. There is a transition between these two, marked by a change of the landform from an essentially flat surface in the east to a very hummocky till unit in the west. The lower B.B.1 surface probably resulted from the erosion and redistribution of the gravels originally contained within the B.B.2 surface. This contention will be discussed further in a later section.

A second major gravel deposit exists in the northern section of the study area (site 11 in Figure 3.5). The morphology of the surface suggests that this is a zone of outwash deposition originating from the "Athabasca Lobe" to the north and west. Several large channel

scars are visible, on aerial photographs, radiating from an area where a series of large, relict, meltwater channels coalesce. Numerous gravel pits are located in this unit and these reveal moderately to well sorted gravels in a sand matrix. Occasional lenses of sand occur within the gravels. Paleoflow directions throughout the vertical sections vary although they remain fairly consistent in the horizontal planes, indicating a general northwest to southeast direction.

Numerous gravel deposits are located along the perimeter of the Red Deer River Valley, approximately 50 to 60 metres above the present floodplain. Exposures show poorly sorted gravels in a medium to coarse sand matrix with some large lenses of fine to medium sand. These deposits vary significantly in their sedimentary composition. In some locations the materials appear to have had a strong fluvial influence with well defined channel fills, very similar to the deposits mentioned above. On the other hand, some of the exposures show very poor sorting, with gravel set in a poorly bedded matrix consisting of sand, silt and clay. These deposits are also slightly cemented due to the presence of clays. For a comparison of such exposures see Figure 3.14, A and B.

It is not known whether the series of gravel pits located along the margin of the Red Deer River Valley represent the remnants of a once more continuous, partial valley fill or if they are isolated deposits of ice contact materials. The variability of the sorting characteristics and the internal sedimentary structures suggest that they were not deposited in the form of a sandur, especially with respect to the more cemented deposits, which show little evidence of fluvial

Figure 3.14 A and B Gravel pits located along the perimeter of the Red Deer River Valley approximately 50 to 60 metres above the present floodplain.

A

B

deposition. Their sedimentary heterogeneity, proximity to the valley wall, as well as the occurrence of smaller features well away from the Red Deer River Valley, suggest that these features reflect localised ice-contact environments.

A further feature associated with the above deposits, bordering the Red Deer River Valley, is a deposit of sand near the town of Sundre (site 12, Figure 3.5). The deposit consists of well sorted, bedded sands and silts. Severe contortion of the beds has occurred and a number of faults and diapiric structures are now exposed (Figure 3.15). This feature is found at an elevation well below those of the gravel deposits mentioned above. Boydell (1970) suggested that this feature is a kame, an interpretation which is accepted here. The high degree of bed distortion probably resulted from the movement of the deposit as a block, possibly in a frozen state. Alternatively it may have been deposited over dead ice. With subsequent meltout of the underlying ice, subsidence may have resulted, causing the contortions.

3.4.6 UPPER RED DEER RIVER TERRACES

A number of terrace surfaces within the Red Deer River Valley were identified. The highest was found in the foothills zone in an exposure bordering the Red Deer River (site 13, Figure 3.5). This deposit is located approximately 20 to 30 metres above the present floodplain level. It consists of approximately one metre of silty clay, light grey in colour, having a columnar structure. Very faint laminations exist below the soil organic root layer. Underlying this

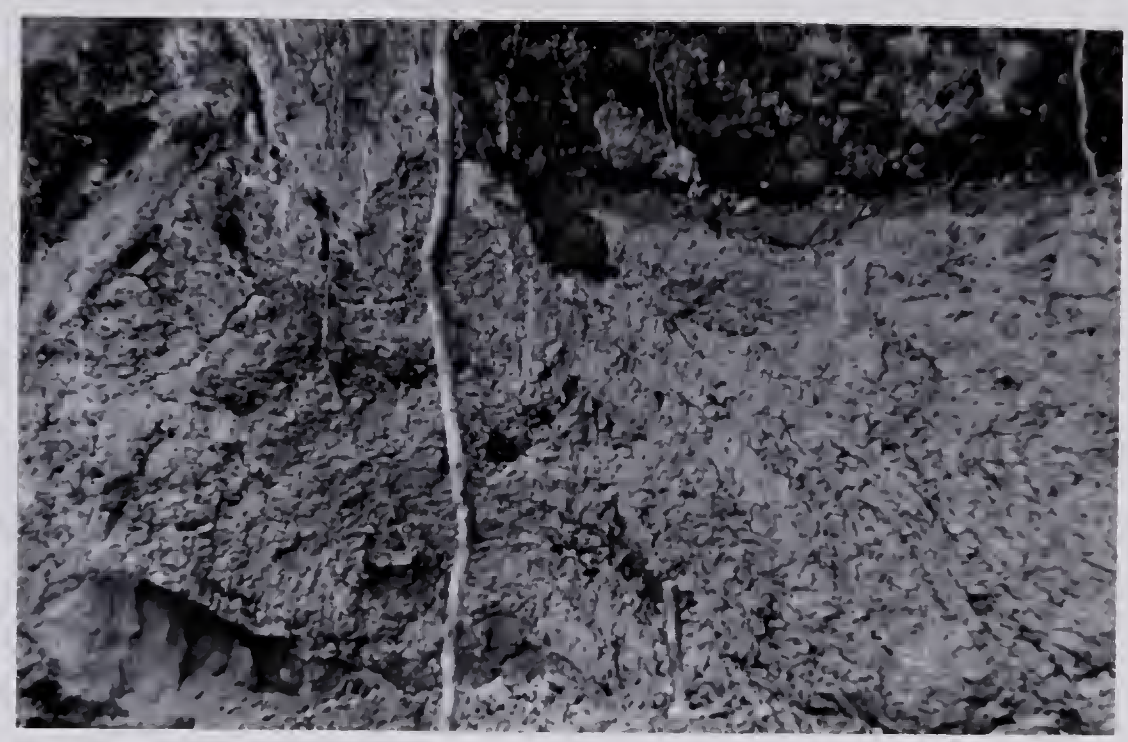


Figure 3.15 Diapiric structure found in kame near Sundre.

is a bed of silty sand interbedded with clay. This unit is approximately 0.4 metres thick. The sand beds range in thickness from 1 to 10 centimetres and show faint parallel laminations. The clays are structureless and have uniform thicknesses of approximately 2 to 3 centimetres. They are succeeded by a layer of silty clay, approximately 15 centimetres thick, in which some parallel bedding is evident. The fine-grained beds are underlain by an unknown thickness of gravel, poorly sorted in a medium to coarse sand matrix. The paleoflow directions of the gravels are consistent with the present direction of flow of the river, that is from west to east. Although the lower portions of the section are partially covered by rubble, it appears that there are two fining upward sequences in the gravel component. The material underlying these gravels is unknown. The deposit is of limited extent and is not found in association with distinctive landform units. There may have been a past linkage, however, between these sediments and an extensive terrace-like form on the south side of the river. Due to limited time and access this possibility was not investigated.

An extensive, irregular terrace located within the valley was traced from the western edge of Bearberry Prairie downstream to the Raven River. The tread is not continuous throughout and is largely absent in the area north of Sundre, where the river valley is relatively narrow, although small isolated remnants were found in this portion of the valley. In addition, a continuation of this surface was found in the William's Creek Valley. The terrace in the upstream section, around William's Creek, appears approximately 2 to 3 kilometres up-

stream from the confluence with the Red Deer River. It is paired, with the eastern remnant abutting against one of the bedrock ridges, and is cut from bedrock. On the western side of the valley the terrace appears to be composed of till which has a high clay content with some Cordilleran gravels. The terrace disappears at the junction of the William's Creek Valley with the Red Deer River Valley where associations of local terraces from the two valleys are intermixed (Figure 3.16 A and B).

On the basis of relative height relationships the B.B.2 surface is probably a continuation of this terrace, although, no remnants of the surface were found between William's Creek and the Bearberry Prairie. This may be because extensive mass movements have occurred in this area, removing most unconsolidated materials from the slopes.

Small remnants of the terrace are located downstream from the Bearberry Prairie on the west side of the valley. Around the mouths of the large gullies, described earlier, they occur as small isolated forms, sometimes extending into the gullies from the main valley. No exposures of the constituent materials were found. At the junction of the Red Deer River Valley and the Eagle Creek Valley, a large, continuous remnant of this terrace exists. In this locality it occurs mainly on the east side of the valley, although smaller remnants persist on the west side downstream from Garrington Bridge. The James River Valley also contains paired terrace treads at comparable heights. The materials composing the terrace in that valley

Figure 3.16 A Oblique, aerial photograph of William's Creek Valley. Note the terrace-like feature around perimeter.

Figure 3.16 B Terrestrial photograph from William's Creek Valley. Note the terrace-like form seen in A in left hand portion of the photograph.

vary greatly, from till, to till interbedded with rhythmically deposited sediments. These sediments are apparently overlain by overbank alluvium consisting of horizontally bedded fine sands, silts and clays. The surface morphology is hummocky. A number of the small relict channels, described in section 3.4.4, occurring at varying elevations, are located on this surface. At the confluence of the James River, and in the northernmost part of the field area, the surface has a cover of eolian sand.

In the foothills zone several small terrace remnants were found. In the area around William's Creek a number of tread levels were noted, although it is was not determined whether these are paired or unpaired. They are composed of gravels with little obvious overbank material, but further downstream the amount of overbank materials increases. On the lowest terrace level this layer of alluvial fines is one metre in thickness. The upper 0.5 metres consists of fine sand and silt with no sedimentary structures. Below this, 0.5 metres of somewhat similar materials occurs but this layer is characterised by very fine parallel laminations. No coarse materials were found in this deposit which was underlain by poorly sorted gravels in a medium to coarse sand matrix (Figure 3.17).

The less extensive terrace levels, at higher elevations than the lower terrace mentioned above, do not extend beyond the foothills area. If they formerly existed downstream of the Bearberry Prairie these have since been destroyed by incision and mass movements. The highest of these remnants, near Williams Creek, is located on stabilized

Figure 3.17 Exposure of lower terrace sediments in the foothills area.

colluvial slopes. Very minor breaks in slope exist on which are found Cordilleran gravels. As mentioned above, these slopes have undergone considerable mass movement so that the terrace treads, where they still remain, have been displaced to varying degrees.

A small gravel pit was found on a tread approximately 15 metres above the present river floodplain (site 14, Figure 3.5). Exposures showed moderately sorted gravels in a sand matrix with incorporated lenses of well sorted fine gravels. A thin layer of fine material occurs over these gravels. This bed contains several larger clasts of both local bedrock and Cordilleran materials. This upper bed is thought to have resulted from colluvial processes. The angular sandstone clasts are not present in the lower alluvial beds to the degree found in the upper bed. Second, the overbank deposits found in other exposures do not display these clasts. The third reason for this interpretation is the proximity of the site to the colluvial slopes forming the valley side. The lowest terrace is one which extends throughout much of the study reach and it includes the B.B.1 surface. The surface morphology of this is convex-up in cross section (Outhet, 1977) with a large number of channel scars. A gravel pit near the western edge of the Bearberry Prairie (site 15, Figure 3.5) shows poorly sorted, medium to fine gravels in a silty sand matrix. The gravel imbrication suggests paleoflow directions from the southwest and this material is mainly of Cordilleran provenance (Figure 3.18). A gravel pit (site 16, Figure 3.5), located near the town of Sundre, showed an extensive section of cross bedded sands and gravels. Quarrying activities have exposed gravels to a depth of 10 metres but have been

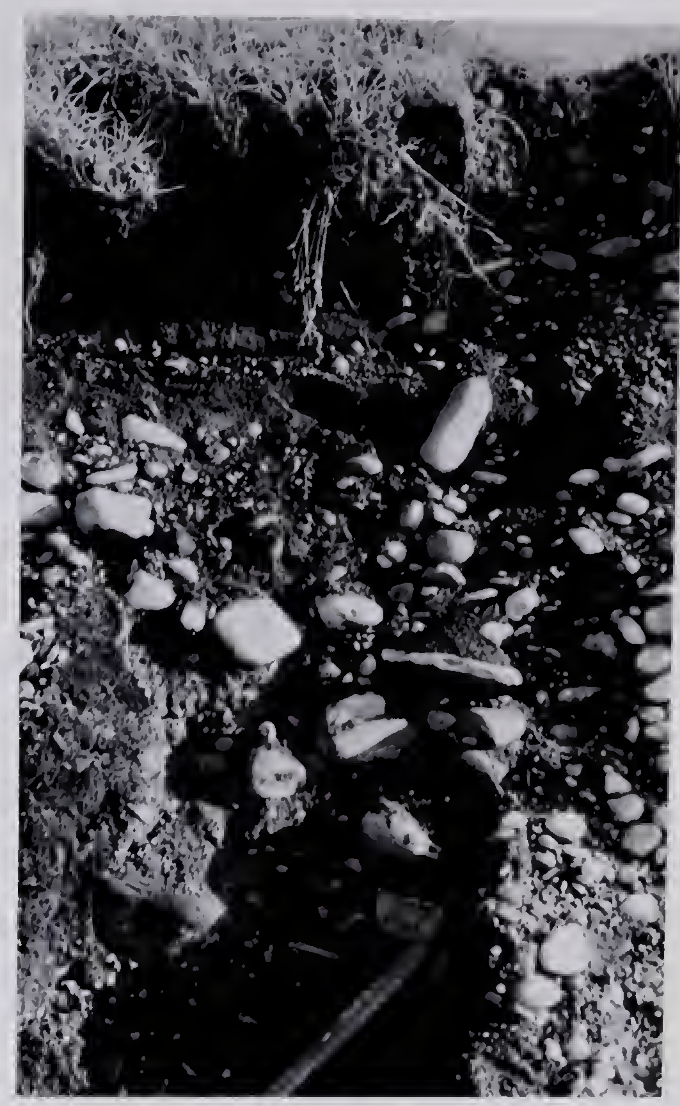


Figure 3.18 Small exposure of gravel on Bearberry Prairie.

halted at this depth by a layer of silt and clay, the extent and depth of which is unknown. The B.B.1 surface is a relatively continuous tread merging into the lower Red Deer River terrace both up and downstream. The segment of the terrace downstream from the Bearberry Prairie is paired and well preserved throughout the reach. It is similar in sedimentary characteristics to the portion occurring upstream in the foothills, with poorly sorted gravels in a medium to coarse sand matrix. A thin cover of overbank fines overlies the gravel.

3.4.7 LACUSTRINE DEPOSITS

An extensive deposit of lacustrine materials exists at site 17, Figure 3.5, bordering the Bearberry Prairie. Several large exposures located west of Fallentimber Creek reveal approximately one metre of very fine, laminated, silty sand overlying an unknown depth of rhythmically deposited sediments. These rhythmites are characterised by apparent "winter" and "summer" beds of variable thickness. The "summer" beds are coarse textured, consisting of silty sand and varying in thickness from 6 to 25 centimetres. Some fine parallel laminations were noted in these beds. The "winter" beds are composed predominantly of clays and possibly contain some silts. The beds are much less variable in thickness than the coarser deposits described above, ranging from 2 to 3 centimetres (Figure 3.19 A). The deposits have pronounced deformation of their beds. This deformation has not destroyed the bedding, as can be seen in Figure 3.19 B, but the beds

Figure 3.19 A Exposure of lacustrine sediments along southern boundary of Bearberry Prairie.

Figure 3.19 B Close up of the exposure shown in A.

have become distinctly contorted. This contortion increases towards the top of the deposit. The deformation found in this deposit is very similar to that noted in the lacustrine sediments mantling parts of the hummocky moraine (Figure 3.8). The surface morphology shows a number of shallow depressions and the unit most probably had a former small thickness of buried ice. Melt-out of this presumed ice probably led to the development of the depressions. The subsidence may represent differential compaction of the sediments. Another possibility is that periglacial processes distorted the sediments after the draining of the lake.

Further upstream a much smaller exposure (site 18, Figure 3.5) reveals a thin deposit of fines and rhythmites approximately two metres thick. The thickness of the individual beds is much more variable than in the previously mentioned section. This deposit overlies a diamicton which has a high gravel content, of Cordilleran origins, in a sandy clay matrix (Figure 3.20). The material is probably of glacial origin but because it may have been relocated by slope processes it cannot be confidently classified as a till. The lacustrine materials decrease progressively in thickness towards the west. At site 19, Figure 3.5, an exposure revealed no lacustrine materials between the alluvial overbank materials and the diamicton.

A second area of lacustrine deposition is within the highlands area to the north of the James River and east of Garrington Bridge (site 20, Figure 3.5). A related, broad, shallow depression is flanked by very subtle breaks in slope. The underlying materials are heterogeneous in composition, varying from sands to clays over very short

Figure 3.20 Stratigraphic section showing overbank fines over a small thickness of lacustrine sediments. This is underlain by a diamicton,

distances. Unfortunately these sediments are poorly exposed so the local stratigraphy could not be ascertained. On the east side of the Red Deer River the shallow channels, described in section 3.4.4, are interpreted as overflow channels from this area of ponding. The relatively shallow depth of both the basin and the channels indicate that the ponding on the surface was of a very short-term nature. This idea will be elaborated within the perspective of the regional pattern of glacial retreat in the discussion section of the present chapter.

3.4.8 EOLIAN AND COLLUVIAL DEPOSITS, HOLOCENE ALLUVIUM AND RECENT DRAINAGE

Deposits of eolian cover sands and dunes exist in the northernmost parts of the study area, on both sides of the Red Deer River Valley. These deposits consist of well sorted, fine to medium sands, with gravels almost entirely absent. The dunes have been modified so that characteristic dune shapes usually have not been well preserved. The long axes of the few distinctive forms have a northwest to southeast orientation (Figure 3.21), and partly transgress the northern outwash gravel deposit described earlier. Dunes also occur on the east side of the river near the confluence with the James River, and north from the confluence with the Raven River. The dunes in these two areas do not display a preferred orientation but, rather, exist as a series of hummocks.

A second type of eolian deposit is exposed at occasional sites, as in the Eagle Creek Spillway (site 21, Figure 3.5). This deposit is composed of a fine structureless silt and its thickness in one

Figure 3.21 Eolian cover sands and dunes located in the northeast section of the study area.

exposure is approximately 0.6 metres. This is separated from underlying till by a stone line 0.3 metres thick. The stone line shows signs of marked post depositional disturbance probably by frost action. A number of the stones have their long axes aligned vertically while others have migrated upwards into the lower portion of the silt unit. The deposit is interpreted as loess (Rains, pers. comm., 1978).

Mass movement processes have been responsible for a variety of landforms in the area. Deep seated mass movements are now largely inactive although in some areas superficial movements, such as soil creep, are persistent. Most of the major, rapid mass movement sites are located in the foothills, especially on slopes facing south along the river valley. In some road cuts (for example site 22, Figure 3.5) colluvial debris was found mixed with fluvially deposited materials. As shown by Figure 3.22, the top of the exposure has a large amount of locally derived debris which is angular and poorly sorted. The lower units are better sorted and contain subrounded clasts, mainly of Cordilleran origin. There is a gradual decrease in the amount of locally derived material towards the base of the exposure, which suggests that the colluvium here largely post-dates the alluvial deposition.

A number of inactive mass movement features occur downstream from the foothills. These encompass the hill slopes near Sundre and forms in the large gullies described earlier. Some small, localised zones also occur further downstream. The morphological characteristics of these are relatively gentle slopes which have an undulating appearance. They are composed of diamictites, most likely relocated tills, and are largely the result of slope failures triggered by basal

sapping of streams and rivers.

Active mass movement occurs in a number of places, the most spectacular of which is an area of particularly active mud flows (site 23, Figure 3.5 and Figure 3.23) associated with the contact between the Belly River and Edmonton Formations near the foothills-Bearberry Prairie boundary. Soil creep also occurs on nearby slopes which are gentler than in the area prone to mudflow activity. The soils on these slopes are shallow, with bedrock outcropping in several places. The vegetation cover is sparse with grasses being dominant. Surficial deposits may have been deposited here in the past but, if so, have been largely removed. In some areas, though, evidence of previously existing glacial or glaciofluvial deposits were found in the form of small, well rounded Cordilleran gravels. A second zone of active mass movement occurs along the bank of the Red Deer River at site 17, Figure 3.5. The lacustrine materials of this area are presently undergoing substantial failure along the river's edge. This will be discussed to a greater degree in the section on bank materials, Chapter Four.

The floodplain of the upper Red Deer River consists of well sorted gravels in a fine to medium sand matrix. Overbank alluvial materials in most of the areas are nonexistent. The floodplain surface has numerous, quasi-relict, channel scars as well as active backwater channels. The vegetation cover varies greatly. In the foothills, the floodplain is vegetated mainly with spruce and mature aspen. The active channel surfaces are for the most part nonvegetated with the exception of perennial grasses. Downstream of the foothills there

Figure 3.22 Alluvial deposits overlain by colluvial materials.

Figure 3.23 Active, superficial mass movements and gullying in the foothills.

is a vegetation progression, towards the active channel from the lower terrace and across the floodplain, from spruce to immature aspen poplar and willow.

Other smaller areas of Holocene alluvium are associated with tributaries of the Red Deer River. The main one is the James River floodplain which is basically that of a single-channel river. The sedimentary characteristics of this floodplain are partly similar to that of the Red Deer River in that gravels predominate in a sand matrix. The large gullies, previously described, have a small terrace remnant consisting of silts and clays with very little gravel observed. The small streams presently occupying them are contained within floodplains consisting of silts and clays.

Holocene drainage systems have in places re-occupied former preglacial valleys or valleys of glacial and fluvioglacial origin. One notable exception is Fallentimber Creek which joins the Red Deer River upstream from Sundre. This stream has incised into a lacustrine basin. William's Creek and a tributary, Brown Creek, are both underfit and occupy what are apparently two old meltwater channels and associated lacustrine basins. Bearberry Creek, which forms the northern boundary of the Bearberry Prairie, has formed partially in a glacial spillway channel (Outhet, 1977). The James River, which is by far the most important tributary to the upper Red Deer River, appears to have re-occupied a preglacial valley. This conclusion is based on the till deposits found within the valley proper, below the high terrace tread, which is an extension of the one described for the Red Deer River Valley. The remaining tributaries are fairly insignificant in terms

of their geomorphic relationship to the Red Deer River.

Large areas, such as the Bearberry Prairie and the northernmost gravel deposit appear to have internal drainage. This is evident from the general lack of surface drainage features or zones of ponding. The floors of some of the meltwater channels and spillways have poor drainage which is expressed by the development of muskeg.

3.5 DISCUSSION

Boydell (1970, 1978) attempted to reconstruct the late Pleistocene history of part of the study area. He interpreted glacial advances from three potential sources; two Cordilleran, Rocky Mountain sources and one of continental origin. The Rocky Mountain ice masses were identified as, first, the "Saskatchewan Lobe" and second, a glacier contained within the upper Red Deer River Valley. The continental, Laurentide ice sheet comprised the third source. These should necessarily have resulted in the deposition of a minimum of three substantially different tills but three distinctive till units were not found during this study, nor were they found by Boydell (1970). His differentiation of the tills, on the basis mainly of particle sizes, was found to be unsatisfactory due to the variation of parent materials in the source areas of the tills. The textural variations of the tills over a short distance were quite noticeable even without laboratory analyses. Second, the stratigraphic evidence does not necessarily support this multiple glaciation theory. The sequence noted earlier, that is till-rhythmites-till, occurs over a very restricted area

and could easily reflect varying depositional processes of one glaciation rather than multiple glaciations. In addition to this the Canadian Shield granites found in the tills do not in themselves necessarily indicate the presence of Continental tills but rather a possible intermixing of various older deposits. This point cannot be conclusively resolved in this study, as it requires more detailed analysis of till properties.

Boydell (1970) suggested that the glaciations did not occur synchronously in the area, but that the ice originating from the Red Deer River Valley had retreated prior to the advances of the "Saskatchewan Lobe" and the Continental ice sheet. This is a fair speculation in that the response times of the glaciers probably varied considerably, with the smaller Red Deer Valley ice mass being effected much more readily by climatic amelioration or deterioration than the larger Athabasca-Saskatchewan and Continental ice masses. Second, the source areas of the three glaciers were in sufficiently different environments that the climatic variables controlling their activity would vary. This is relevant to the sequence of events interpreted for the Red Deer River Valley during deglaciation.

The occurrence of the till deposits within the upper Red Deer Valley, suggests that this section of the valley is "preglacial". in terms of existing prior to the last glacial advance. The original valley, however, was subjected to major modifications by the ice mass. The advancing ice, from the north and west, occupied the valley as well as its surroundings. The gravel deposits, interpreted as kames, found along the perimeter of the valley substantiate this conclusion.

These deposits were probably formed in part by the reworking of glacially deposited materials, in contact with the valley ice, by glacial meltwater.

Shaw (in press) suggested a model which stressed meltout of debris from the bottom of a glacier as an important mechanism of glacial deposition. He stated that:

> "...with the development of a surface debris cover and subsequent melting, till is released by basal meltout...The poorly attenuated facies undergoes little deformation so metamorphic foliation of debris and ice, and clast orientation are poorly developed"
> (Shaw, in press)

Shaw and Archer (1978) suggested that this meltout process produced, in the Okanagan Valley, British Columbia, extreme variations of depositional sequences over short distances. They described sands and gravels surrounded by tills, as well as tills in contact with each other over a distance of 100 metres. They attributed variations such as these to the melting out of basal debris and the superimposition of supraglacial and englacial material.

It is suggested, therefore, that in the case of the upper Red Deer River Valley, the stratigraphic sequences occurring at sites 1, 2, and 3 (Figure 3.5) may be explained by this model. A number of reasons for its applicability may be advanced. The first is the proximity of the exposures to one another, and the substantial variation of their exposed stratigraphic sequences. The second reason lies in the compaction, or induration, of the glacial deposits. The lower till is more compacted than is the case for the upper one. This suggests that the lower till unit resulted from basal depositional

processes whereas the upper unit resulted form supra-, en- or subglacial meltout. This aspect of the glacial history of the upper Red Deer River Valley is far from understood.

The channels occurring in the northern sections of the field area may be divided into two groups; a) meltwater channels incised into bedrock and b) spillways and overflow channels incised into unconsolidated materials. The meltwater channels have all developed approximately parallel to each other but they trend perpendicularly to the regional slope. This suggests that some obstruction prevented them from developing a northeast trend. Such an obstruction may have been a stagnating ice lobe which produced a series of channels, not contemporaneously, but in a chronological order, decreasing in age from south to north. Once initial development occurred incision continued until the local meltwater sources, stagnating ice, were depleted. A number of terrace treads occur in two of these channels. These remnants may represent alterations of the local base level caused, at least partially, by the melting of ice in the upper Red Deer River Valley and partially by changes of Red Deer River incision rates. As mentioned earlier, these terrace surfaces only occur in a restricted number of channels and do not appear in channels located north of the study area.

On the eastern side of the river smaller and more subtle channels are incised into unconsolidated deposits. These channels are located at a number of different elevations and may be overflow channels associated with ponding of meltwater which had collected in the Red Deer River Valley, with the differing elevations corresponding to the

progressive melting of the ice contained within the valley.

South of Sundre lacustrine materials mark the location of former glacial Lake Sundre (Boydell, 1970). The relationship of this lake to the deglaciation events is not fully understood but it is believed that it was formed during the latter stages of deglaciation when ice, acting as a drainage barrier, was present in the Red Deer River Valley to the north. The stratigraphy of the deposits, and associated landforms, indicate that the lake was most probably formed during deglaciation. Shaw (1977, p. 229) stated that "the upwards decrease in thickness of varved units is characteristic of classical glacial lake sediments deposited during retreat". The upward decreasing thickness of the lacustrine beds is displayed in Figure 3.19 A. A related point is that the $B.B._2$ surface is probably a remnant of a deltaic deposit, consisting mainly of gravels. The exposure described for site 13 (Figure 3.5) is considered to be a western extension of this unit. The fines overlying the gravels are probably lacustrine materials, indicative of a transgression of the lake over the delta. Boydell (1970) stated that the $B.B._2$ surface was deposited during the initial stages of the development of Lake Sundre, that is, well before the final deglaciation of the area, and that a subsequent advance of Laurentide ice overrode the deposits. No evidence could be found by the author to support this contention. The morphologies of the $B.B._2$ surface and the nearby lacustrine deposits suggest that little post depositional disturbance, other than the effects of minor subsidence, has taken place. Outhet (1977), on the other hand, suggested that the $B.B._2$ surface is a kame and that the lower $B.B._1$ surface is an

outwash fan. Based on evidence previously discussed, and the fact that the deposit is located topographically below the lacustrine deposits, this conclusion must be questioned.

Speculation as to the late glacial, or deglaciation, history of the upper Red Deer River Valley may therefore be based on the evidence presented above. Deposition of glaciofluvial and glaciolacustrine materials in the form of the B.B.2 surface and adjacent lake deposits occurred by meltwater originating mainly from the Cordillera and foothills but also from ice located to the north and east. Ponding was caused by ice stagnating in the Red Deer River Valley downstream from Sundre. Draining of Lake Sundre occurred primarily as the ice within the valley progressively wasted. The lake would have formed the local base level for the Red Deer River flowing through the foothills. Subsequent drainage development, possibly in discrete stages, would have stimulated episodes of incision and periods of relative stability to create the series of terraces found in the foothills, but not located downstream of Sundre. Incision and reworking of the Bearberry Prairie gravels, to form the B.B.1 surface and associated terrace treads, occurred after Lake Sundre was finally drained.

The large gullies located to the north of Sundre were probably formed during the final stages of deglaciation of the area. It is not known whether drainage lines existed there prior to the last, major, glacial advance. If so, though, they would have experienced considerable modifications during the glacial retreat. The gully systems extend eastward from the hummocky moraine and do not display drainage areas proportional to their sizes. They grade down to the level of

the valley fill in the Red Deer River Valley, as indicated by the continuation of the small terrace remnants into the main valley. The exact history of their formation, however, has been obscured by mass movements along the valley sides.

The northernmost deposit of gravels, in the form of an outwash fan or delta, was formed by material and meltwater from the northwest. This deposit grades down to the valley fill and lacustrine materials previously mentioned. The eolian materials covering these gravels are probably the deflation fines originally contained within the gravel and lacustrine deposits.

A general, relative, chronology of the deglaciation of the region, based on the material presented in this chapter, is formulated as follows. The first areas to become deglaciated were the highlands surrounding the Red Deer River Valley. After the highlands were exposed, the kames located along the sides of the valley, developed. It was also during this phase that ponding occurred north of the James River on the highland surface. Progressive melting of the ice contained within the main valley, below the highlands level, resulted in ponding of water within the valley proper, causing the formation of the succession of overflow channels to the north and east of Garrington Bridge. Meltwater from the northwest, flowing through the large spillways and the large gullies to the north of Sundre, constituted a major source of the water. Lake development continued until the downvalley drainage obstructions were reduced or disappeared. Because exposures of related sediments are rare to the north of Sundre, it is difficult to determine whether this northern zone of ponding and Glacial Lake

Sundre were one continuous unit or whether they constituted a number of separate bodies of water. Subsequent to the draining of the lake(s) a period of incision caused erosion of the till and lacustrine deposits contained within the valley, distributing these materials, along with abundant gravels contained within the $B.B._2$ deposit, downstream. Relatively minor amounts of gravel from the tills would also have been added to the alluvium. The fines contributed by the lacustrine deposits and the tills were largely washed out of the system resulting in the gravel-rich sedimentary character of the lower terrace. A second period of incision followed culminating in the present flood-plain level.

3.6 CONCLUSIONS

In summary, the gravels occurring in the Red Deer Valley are predominantly of Cordilleran origin, were deposited during the final stages of deglaciation of the area, and were partly redistributed by more recent fluvial processes. Some of these gravels are presently being reworked as part of the active floodplain. The gravels found to the north of Garrington Bridge, and associated with the relict melt-water channels, may have added relatively minor increments of debris to the Red Deer River during its Holocene evolution in this area.

Although the major sources of large clasts have been identified in this chapter the Holocene evolution of the floodplain up to its present day configuration is not clearly understood. Detailed examination of the alluvial stratigraphy and the acquisition of relevant radiocarbon dates are required for a precise explanation of the river's late Quaternary history.

CHAPTER FOUR

BANK MATERIALS AND STABILITY: CHANNEL SEDIMENTS

4.1 INTRODUCTION-BANK MATERIALS AND STABILITY

Bank stability, or the ability of the river banks to withstand erosion from fluvial processes, depends primarily on two principal factors. These are, first, the nature of the materials composing the banks, and second, the type and density of the vegetation growing alongside of the river. After an extensive literature search, little material concerning the erosive properties of gravel, or non-cohesive, banks was found. On the other hand, numerous papers have been published on the characteristics of the finer, cohesive bank materials and their responses to fluvial erosion. Wolman (1959) found that the banks of Watts Branch, Maryland, were most susceptible to fluvial erosion when a combination of thorough wetting of the constituent materials and a high discharge occurred concurrently. This he noted, caused first, a reduction in the cohesion of the bank sediments and second, removal of these, now more easily eroded materials, by the higher flow velocities. He found a similar result with a combination of frost action and simultaneous high flows. The frost action separated the particles and aggregates of material thereby reducing the cohesion factor. Schumm (1960) found that bank resistance varied directly with the amount of silt and clay contained within the sediments; that is, with

the implied cohesiveness of the bank materials, He argued that banks containing little or no fine cohesive material are highly susceptible to erosion, and promote generally wide and shallow river cross sections. On the other hand, those banks which contain large amounts of silt and clay promote narrow and deep channels because of their relatively cohesive materials. The upper size limit for these cohesive materials was drawn at 0.074 mm., 3.8 ϕ, which includes the silt and clay fraction according to the Unified Classification System (McKeague, 1976). Schumm (1963), incorporating data from a number of rivers on the Great Plains, expanded on his previous conclusions by stating that sinuous rivers showed higher silt-clay ratios in the bank materials than those which were braided. Brice (1964) argued that not only is the quantity of cohesive material important in determining bank stability but the distribution of these materials is also of potential significance. He noted that "... silt and clay concentrated in layers gives a different cohesive effect than would the same amount of silt and clay distributed uniformly throughout the bank materials" (Brice, 1964, p. 33). The specific difference was not discussed although presumably banks with cohesive material organised into separate beds would be more resistant to erosion than in the case where they were evenly distributed throughout.

Of further interest is the effect of vegetation on bank stability. Deep rooted, riparian vegetation will have a binding effect on the bank materials. More important, however, is the effect of a vegetative mat, consisting of shallow rooted grass. Smith (1976) noted the effects of a vegetative mat in reducing bank erosion on the

Alexandra River, Alberta. He observed that with a root accumulation of 5 centimetres, a riprap-like cover was developed after a short, initial period of erosion. This vegetative mat, draped over the bank, inhibited more extensive bank erosion but a relatively large amount of bed scour adjacent to the bank occurred. Schumm and Lichty (1963), in an earlier study on the Cimarron River, Kansas, reported that grass roots offered good protection against bank erosion. Brice (1964), noted that the riparian vegetation determined the meandering channel pattern by stabilizing the river banks of the Calamus River, Nebraska. He stated that where swamp vegetation occurred the channel banks were confined and a meandering channel had developed.

Burkham (1972), in his study of the Gila River, Arizona, noted that the effects of trees in the stabilisation of river banks may be significant. He argued that during minor and moderate floods the trees added to the stability of the banks. During the more severe flow events, however, the effect would be reversed. Concentration of flow within the original channel, as well as in previously abandoned channels on the floodplain, led to increased stresses and erosion of bank materials. Second, felled trees and other debris may be caught close to the river bank. This will tend to increase the turbulence in the vicinity of the debris inducing localised scour.

Vegetation not only stabilises the banks but, if allowed to develop on exposed channel surfaces, may result in their stabilisation. Schumm (1977, p. 121), stated that, "If a channel with alternate bars is invaded by vegetation during a long period of low flow and if the alternate bars are stabilised by vegetation, the thalweg will

deepen during the next high flows." This, he noted, may lead to a transition from an initially straight channel to one which is sinuous in nature.

4.2 BANK MATERIALS AND BANK STABILITY OF THE STUDY REACH

Northwest Hydraulic Consultants Ltd., (1976) undertook a survey of bank erosion along the Red Deer River from within the foothills portion of the study area downstream to the city of Red Deer. The preliminary results of their investigation imply that approximately 2,200 acres (880 hectares) of land have been eroded along this reach of the Red Deer River during a 25 year period. It was also noted that the area between Sundre and the Raven River confluence had undergone the greatest apparent erosion rates. More detailed field investigations undertaken for the present study reveal that the materials comprising the banks of the upper Red Deer River may be divided into three main groups; gravel, bedrock, and sand, silt and clay (Figure 4.1). The gravel banks are the most extensive and relate to the Bearberry Prairie, the lower Red Deer River terrace as well as the active floodplain. These deposits contain very little silt and clay but have a matrix of medium to coarse sand (Figure 4.2). The proportions of the materials vary, but the coarser fraction dominates. The small quantities of fines, therefore, make the gravel banks highly susceptible to fluvial erosion by bank cavitation. Because of the nature of these materials the approach outlined by Schumm (1960, 1963) is not applicable. It appears that two basic factors govern the stability of the gravel banks. The first includes vegetation. The floodplain in cer-

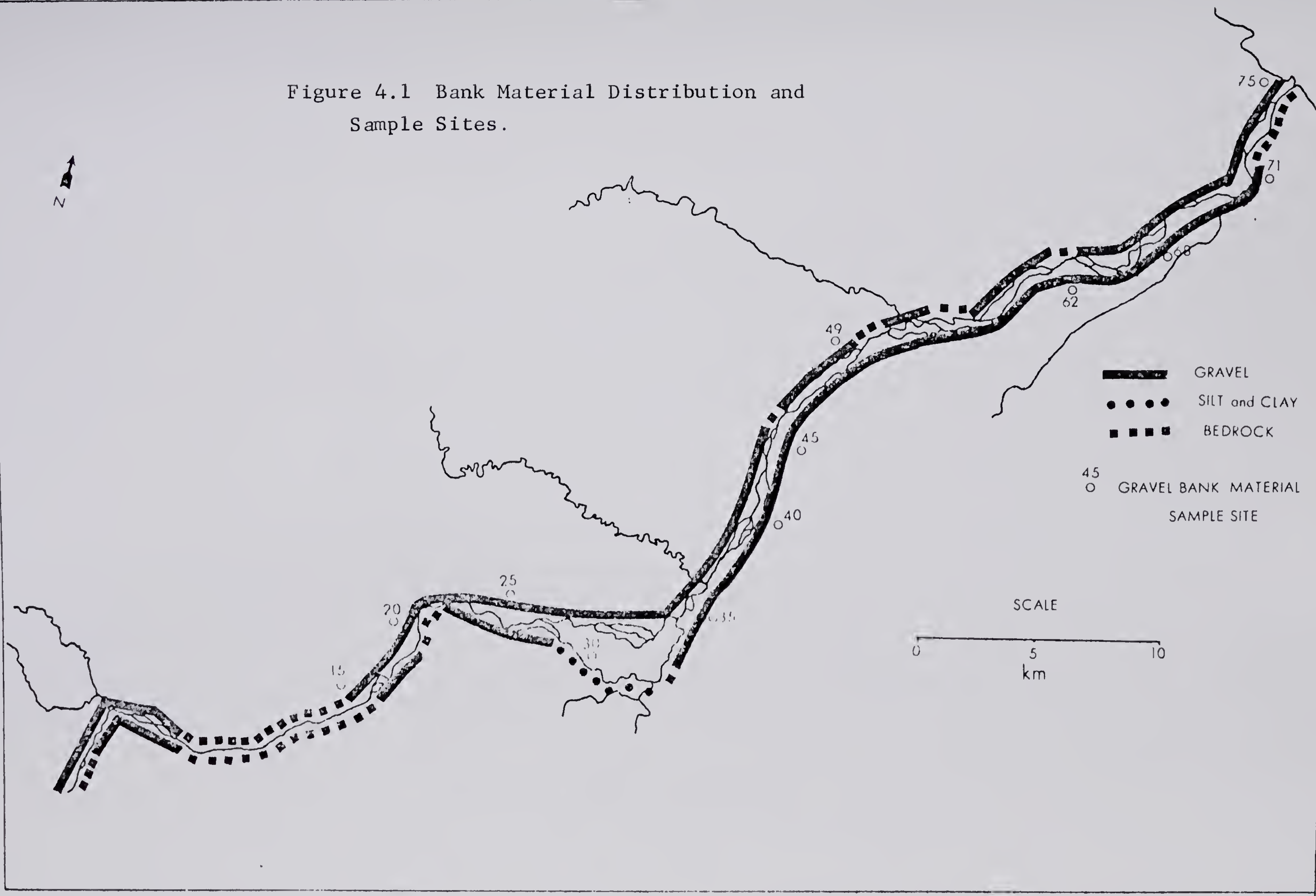
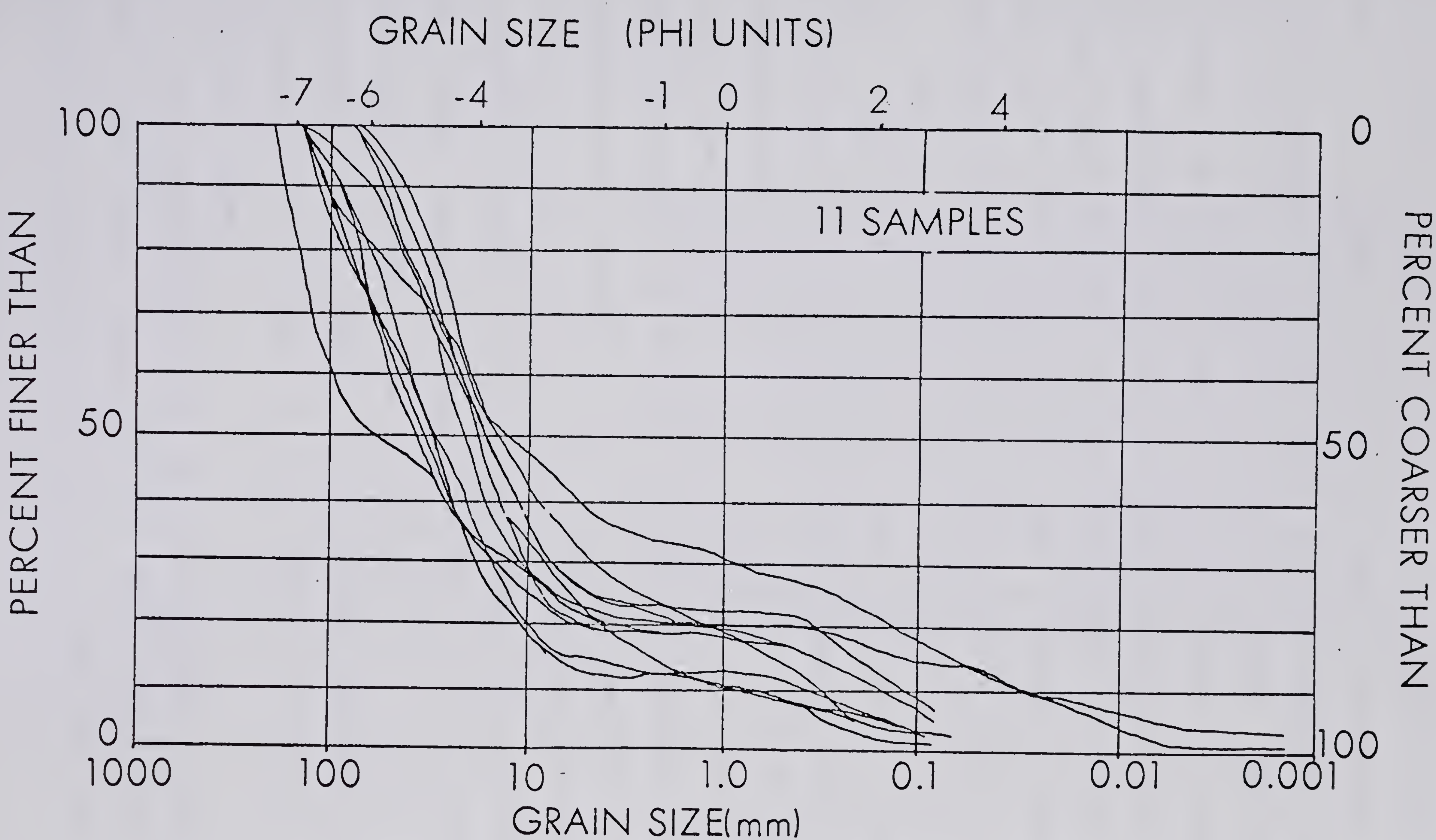


Figure 4.1 Bank Material Distribution and Sample Sites.

Figure 4.2 Grain size distributions of the gravel bank samples.


GRAIN SIZE (PHI UNITS)
-7
-6
-4
-1
0
2
4
11 SAMPLES
PERCENT FINER THAN
100
50
0
PERCENT COARSER THAN
0
50
100
1000
100
10
1.0
0.1
0.01
0.001
GRAIN SIZE(mm)

tain areas has a vegetative mat similar to that described by Smith (1973). It was observed that following an initial period of bank erosion, the vegetation mats would slump down over the sides of the banks, affording some protection against further erosion. The degree to which this has hampered further erosion is not known. The second factor contributing to bank stability is the size range of gravel contained within the banks. Where very large clasts are available these tend to form a pavement along the sides of the banks, thereby restricting further erosion by minor and moderate flows (Figure 4.3).

Bedrock is exposed in a series of outcrops along the study reach. In the foothills it is the most common bank material in that the river has incised below the alluvial terrace deposits into the underlying bedrock (Figure 4.4 A and B). The bedrock forms the most resistant bank material, with respect to fluvial erosion, but the sandstones and siltstones are moderately susceptible to mechanical weathering processes which contribute a potentially significant amount of clasts to the river (Figure 4.5). Relatively low resistance to mechanical weathering processes, as well as some solution of the cementing agents, quickly reduces the locally derived material into its constituent parts (Figure 4.6). In the eastern parts of the study area, the bedrock occurs as a series of isolated outcrops projecting into the path of the river. These have caused substantial deflections of the river channels.

The remaining group of bank materials consist of the lacustrine silts and clays. These are generally found in the area west of the confluence of Fallentimber Creek with the Red Deer River, and

Figure 4.3 Gravel pavement on river bank. Note hammer in foreground for scale.

Figure 4.4 A and B Main channel in the foothills has incised into the local bedrock. Note the occurrence of more resistant beds forming rapids (1) in A.

A

B

Figure 4.5 Clasts of local bedrock, produced by mechanical weathering, being contributed to the Red Deer River.

Figure 4.6 Weathering of locally derived clast on exposed bar surface.

Figure 4.7 Slope failure in lacustrine silt and clay.

form the southern boundary of the Bearberry Prairie (Figure 4.7). The deposit consists entirely of fine sediments, and is lacking in gravels. The cohesiveness of the material theoretically makes it quite resistant to fluvial erosion. However, seasonal groundwater seepage, and some undercutting by the river, have promoted intensive slope failure of the banks. The deposit may therefore be considered fairly susceptible to erosion and is not adequately represented by Schumm's (1960, 1963) silt-clay ratio.

An examination of the size distributions of the gravel bank samples, Figure 4.2, will show the predominant importance of the gravel fraction. The relatively poor cohesiveness of these gravels accentuates bank instability. The banks consisting of bedrock are the most resistant to erosion and appear to be of considerable importance in the development of channel patterns. This topic will be explored in later sections of the thesis.

4.3 INTRODUCTION-CHANNEL MATERIALS

Numerous studies have focused on the nature of channel sediments and the responses of the alluvial materials to fluvial processes and controlling factors. For example, Wolman (1955) stated that there was a direct relationship between the channel slope and the sediment size. He observed, in the Brandywine Creek, Pennsylvania, that the slope of a channel increased with an increase in sediment size. This he related to distance downriver and found that the tractive forces decreased with a decrease in the slope, thereby initiating the deposi-

tion of progressively finer material. Hack (1957) noted that there was a systematic relation between the sediment size and the channel slope. The slope, he argued, was directly proportional to the size of the sediment. He found, however, that variations in the sediment size and sorting were also significant within a channel cross section. The distance from the source of the river did not correlate well with the size of the material. Fahnestock (1969) observed that on the Slims River, Alaska, the lowest channel slopes were associated with the smallest sediment sizes. He found that the slope appeared to be inversely related to the efficiency of sediment transport through the reach. McPherson (1971), in his study of the North Saskatchewan River, Alberta, noted that particle size and sorting correlated much better with channel slope than with distance downriver. The relationships which he presented showed trends consistent with those mentioned above. The relationships were weak, however, with correlation coefficients of 0.42 and 0.43 obtained for size and standard deviation, respectively, against distance. The regressions for size and sorting (standard deviation) against slope showed generally stronger correlations, 0.50 and 0.43 respectively. He attributed these generally weak trends to the input of fresh sediment along the course of the river. These sources were mainly of glacial till and colluvial materials which contributed a heterogeneous mixture of debris to the river.

Knighton (1975) described an exponential decrease in sediment size with distance downriver as well as with declining channel slope. He stated that the sediment size varied directly with the slope, because the steeper slopes were required to allow the transportation

of larger clast sizes. Church and Kellerhals (1978) showed that for the Peace River, in northeastern British Columbia, there existed an exponential decrease in the sediment size downstream. They noted that the slope of the regression line steepened in areas where large tributaries joined the Peace River. In that study, as was the case with McPherson (1971), a significant scatter existed around the regression line. For a number of streams in Kansas, Osterkamp (1978) found that the sorting of the channel materials was inversely related to the channel slope. Poorly sorted channel materials were found on the steeper slopes while the gentler gradients had characteristically better sorted sediments. He concluded that the two main controls of channel gradient were the mean discharge and the mean particle size. His study was conducted with reference to braided streams and, as noted in Chapter One, multiple-channel reaches are commonly associated with relatively steep slopes and poorly sorted material.

A number of workers in the past have discussed the relative importance of abrasion and selective deposition as causes of sediment size decrease downstream. Scott and Gravlee (1968) argued that the variations in sediment size down-river, after the failure of a rock-fill dam on the Rubicon River, California, could be explained almost entirely by selective deposition of sediment during the declining stages of the flood. They noted that the variations in roundness, which were considered an indicator of abrasion, only accounted for a change of less than 5 percent of the total decrease. A general decline of 65 percent of the B axis length occurred after a distance of only 2 kilometres down-river. The 60 percent difference, they argued, must therefore have

resulted from selective deposition and sorting, McPherson (1971), for the North Saskatchewan River, concluded that the decrease in sediment size downriver is "...most probably a consequence of selective stream transport" (McPherson, 1971, p. 76). Shaw and Kellerhals (in press) argued that a strong tendency for the sediment to decrease in size downriver is indicative of an aggrading river. They noted that "...the coarse particles will be progessively buried before 'catching up' with the fines." (Shaw and Kellerhals, in press, p. 23).

4.4 CHANNEL MATERIALS OF THE STUDY REACH

Two types of channel material samples were obtained at selected locations, noted in Figure 4.8. The first type, consisted of a gravel sample and was collected using a central peg and a rope five metres in length, marked at one metre intervals. The sites for the central peg were chosen randomly by throwing a geological hammer on to the exposed channel surface. The rope was then laid on the gravel surface and the particle falling under each mark was collected. The rope was moved after each line of five samples was collected, and samples from twenty lines were taken at each site. The majority of the exposed gravel surfaces were small enough that the ten metre diameter sampling circles, of the method employed, encompassed most of their areas. This technique was used, rather than that suggested by Wolman (1954), because the depth and velocity of the river, even at low stages, precluded complete transect sampling. A photographic technique (Bray, 1971; Bramm, 1977) sometimes used to sample gravels, was also considered

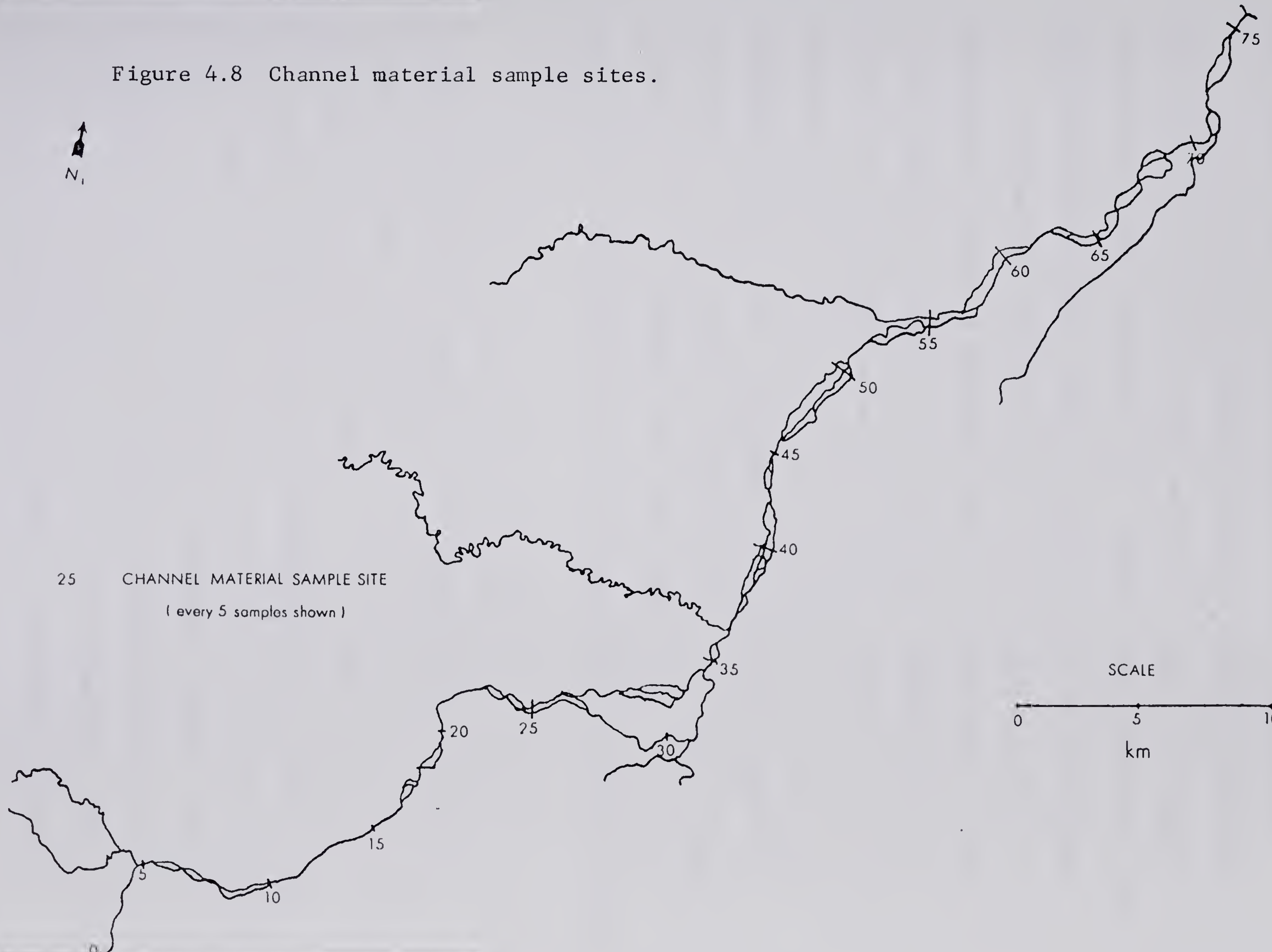


Figure 4.8 Channel material sample sites.

but rejected because of problems imposed by gravel imbrication.

The second type of sample collected was that of bulk form. Here the gravels, as well as their matrix materials, were sampled from the area below the central peg. Samples of both coarse and fine materials thus collected, ranged in weight from 10 to 20 kilograms.

The analysis of the type one, gravel, sample consisted of measurement of the three principal axes (A, B, and C) and recording of the lithological composition of each clast. Roundness characteristics were not recorded for two main reasons. The first, was that the sediments supplied to the river, for the most part, have previously undergone fluvial action. The large clasts were often, therefore, subrounded to rounded before entering the river. The second reason, was that previously McPherson (1971) had shown that particle roundness did not change significantly with distance down the North Saskatchewan River.

Mechanical analysis of the bulk samples, dry sieving and hydrometer analysis, were carried out by Materials Testing Laboratory, Alberta Environment, to provide grain size distributions. These distributions are illustrated in Figures 4.9 and 4.10. The variation in the shapes of these curves is not substantial over a total distance of 75 kilometres. Minor variations do occur, but they may be explained by the variations in the depositional environments sampled. Figure 4.10 indicates that there is no apparent systematic variation in the sediment size distributions of the bulk samples downstream. The curves also show that the gravel fraction normally constitutes between 65

Figure 4.9 Grain size distributions for all of the channel samples.

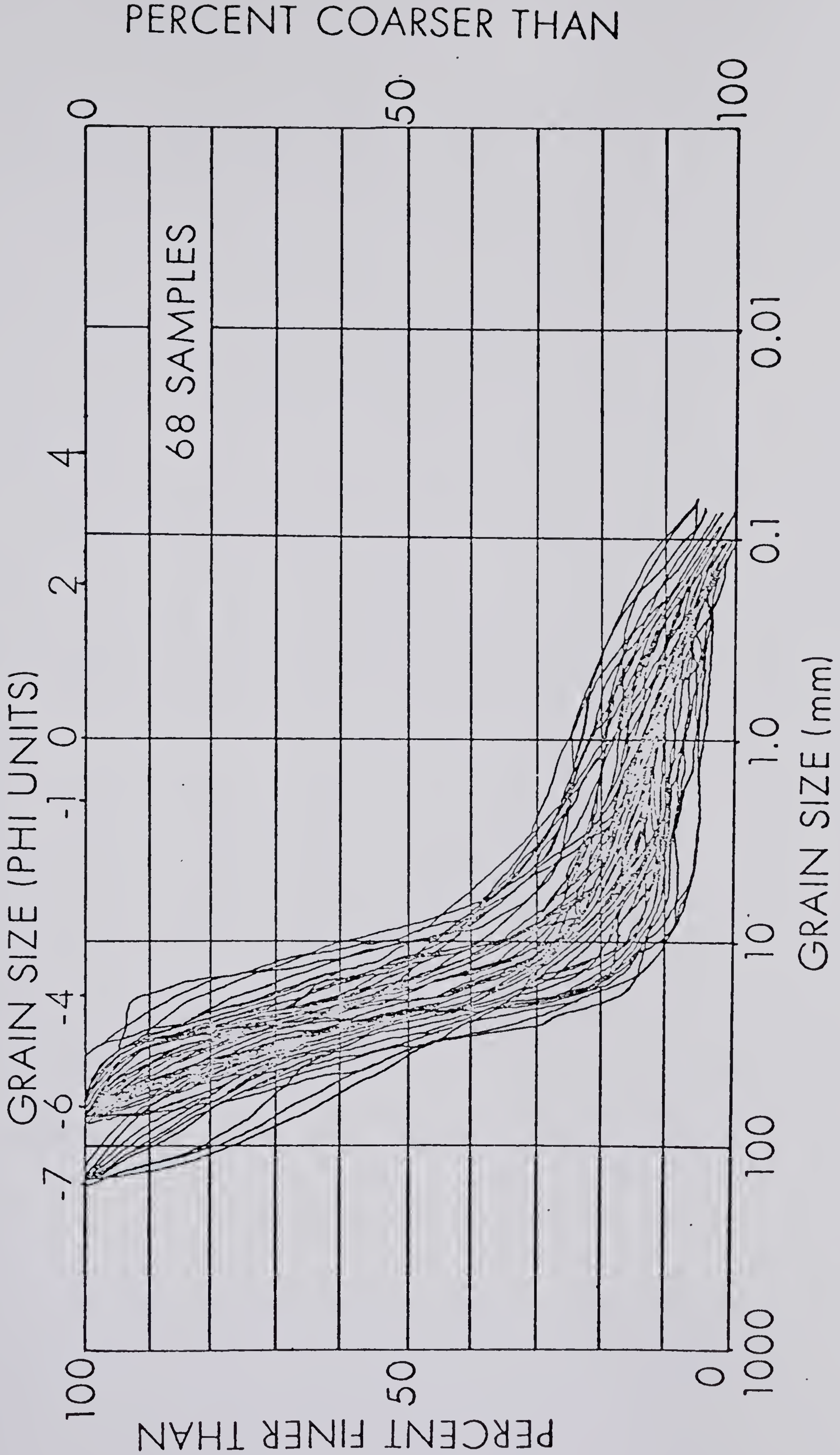
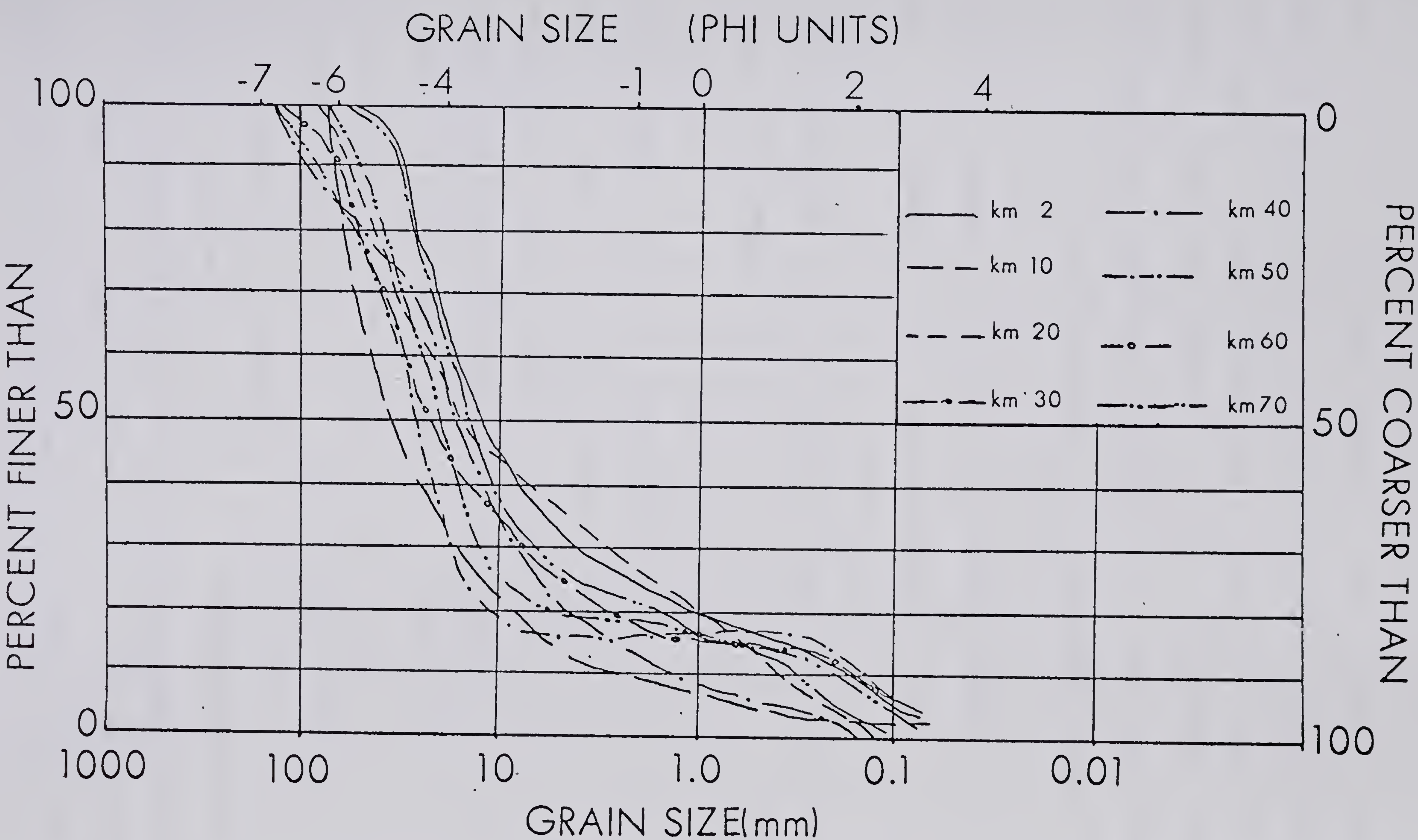

PERCENT COARSER THAN
0
50
100
68 SAMPLES
GRAIN SIZE (PHI UNITS)
-7
-6
-4
-1
0
2
4
GRAIN SIZE (mm)
1000
100
10
1.0
0.1
0.01
100
50
0
PERCENT FINER THAN

Figure 4.10 Grain size distributions for selected channel samples.


GRAIN SIZE (PHI UNITS)
-7 -6 -4 -1 0 2 4
100
50
0
PERCENT FINER THAN
0
50
100
PERCENT COARSER THAN
km 2
km 10
km 20
km 30
km 40
km 50
km 60
km 70
1000 100 10 1.0 0.1 0.01
GRAIN SIZE(mm)

and 95 percent of the channel materials. These gravels, as previously discussed (Chapter Three) have been derived from a number of sources, including till, located predominantly upstream of the study area, and reworked fluvial and glaciofluvial deposits. Local bedrock fragments also constitute a significant proportion of the total gravel sample.

It was thus decided that the type one, gravel, samples should be studied in greater depth, insofar as these were likely to provide qualitative insights as to the role of bedload transport in the study reach. The statistics used to describe the gravel sample size distributions were the mean and standard deviation of the B axis. Inman (1952) recommended the use of the mean, rather than the median (D_{50}), because its derivation incorporates not only the central portions of the sample but also the extremities of the distribution. Folk (1968) suggested that the standard deviation is an appropriate measure of the degree of sediment sorting. McPherson (1971) found that the skewness of sample distributions did not vary significantly downstream in the North Saskatchewan River. The skewness, he noted, is indicative of flow velocity fluctuations. Because velocity fluctuations and distributions vary considerably within a channel cross-section, a realistic skewness value for a specific channel cross-section can only be obtained by the integration of a large number of samples taken across the channel section. Comparisons of skewness characteristics are thus inappropriate for the limited sampling program employed here.

Figure 4.11 shows the percentage of quartzite and sandstone plotted against distance downstream from the western limit of the

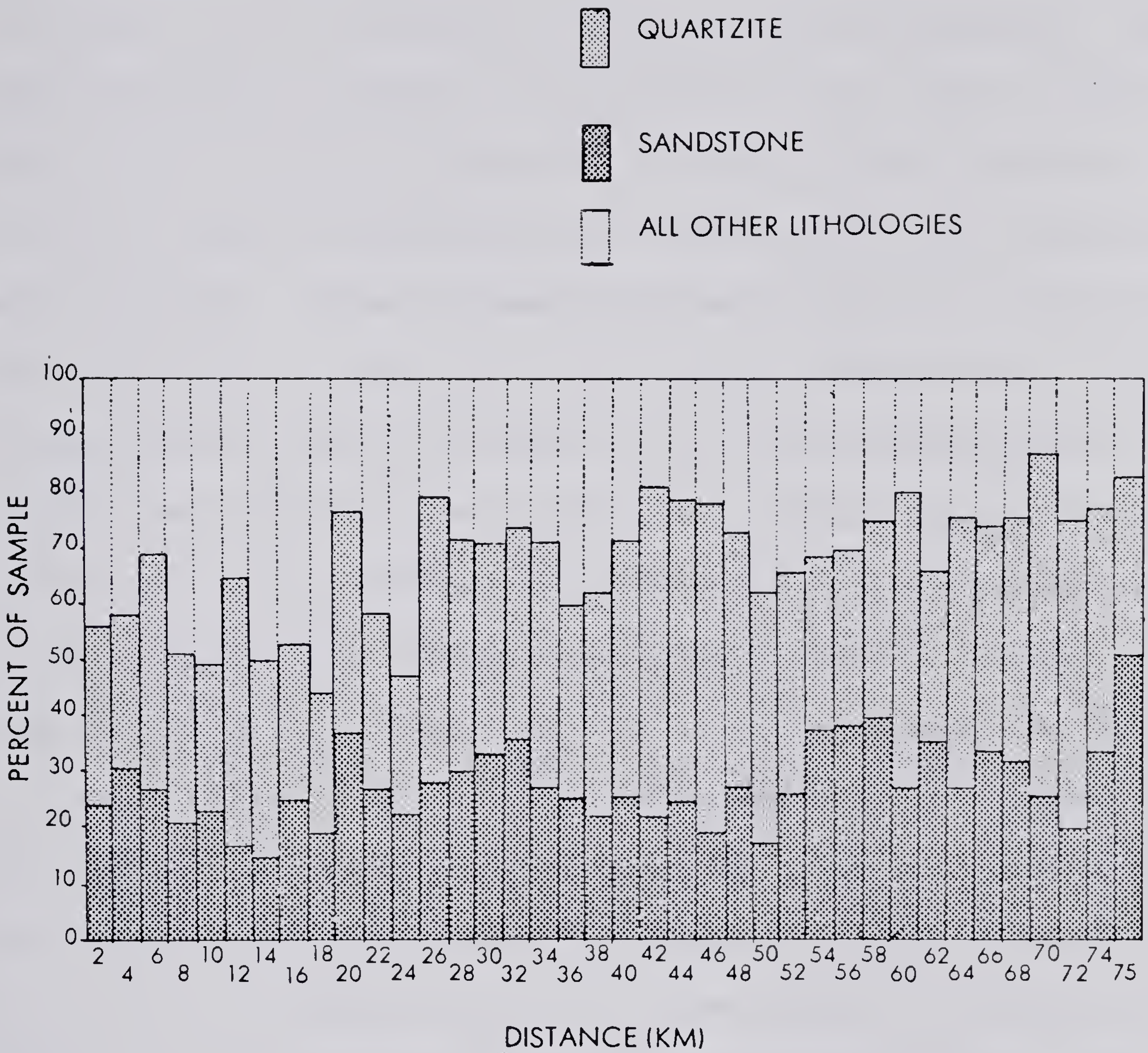


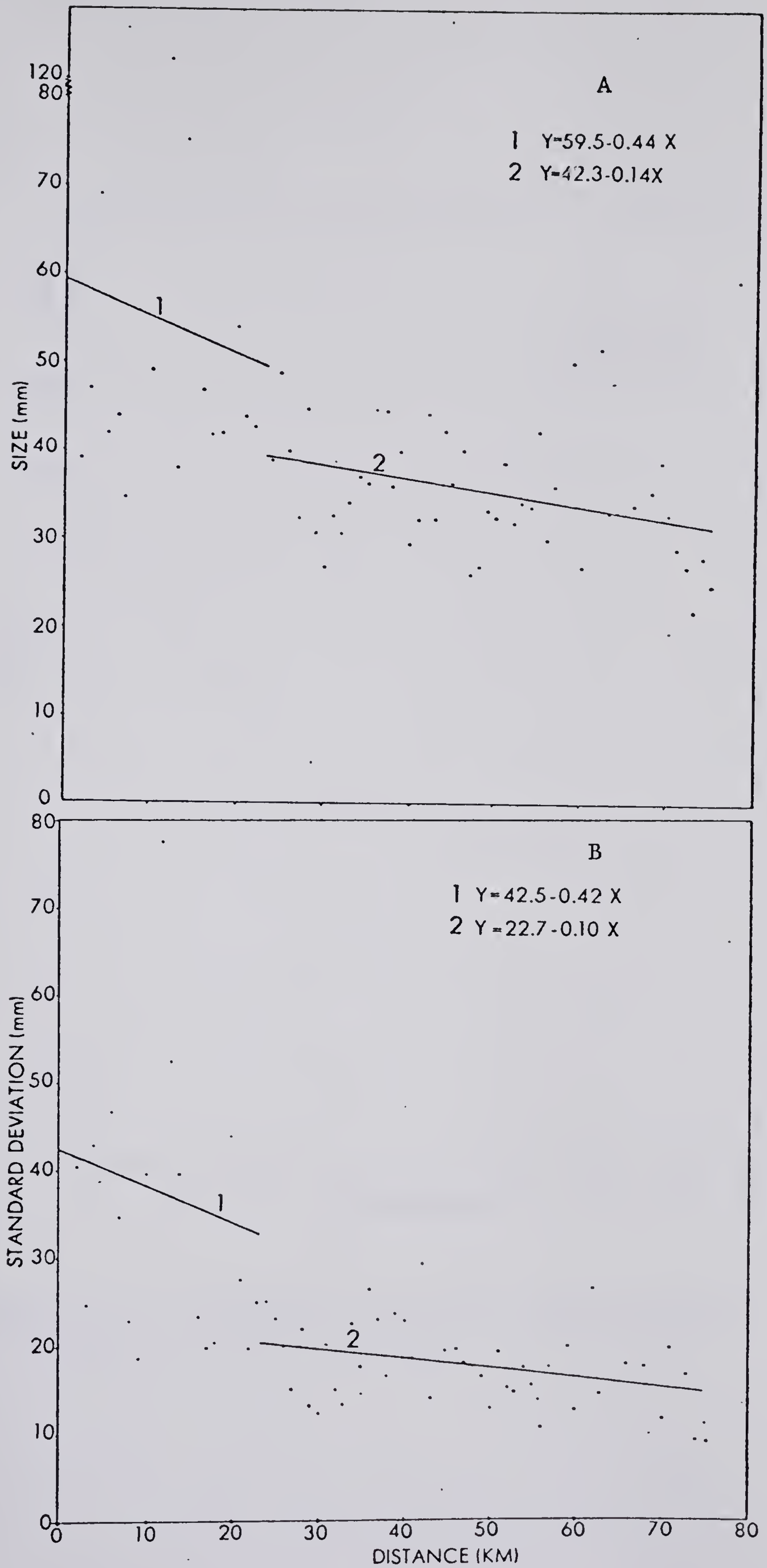
Figure 4.11 Lithologic composition of channel samples.

study area. Quartzite was chosen as the main representative of Cordilleran materials. The other major lithology from that provenance are the carbonates, including limestone and dolomite, which were not differentiated in this study. The quartzites dominated the samples and tended to be more resistant to erosion than the carbonates. Sandstone was chosen to represent the materials of local origin. This lithology is by far the most abundant of the locally derived materials but minor amounts of siltstones, clay ironstone, chert and conglomerates were also noted. The main trends indicated by Figure 4.11 are as follows. First, in the foothills a weak pattern is discernable. Contributions from both the local bedrock as well as the unconsolidated alluvial and glaciofluvial materials occur. Further downstream, the trends become more pronounced. Fluctuations over long distances occur in the relative percentages of the two main lithologies recorded. Large quantities of sandstone are found in the sub-reaches from km 20 to km 34 and km 54 to km 75, with minor fluctuations occurring throughout. The quartzite portion of the sample has a number of peaks, including a sub-reach from km 40 to km 54, as well as a minor one from km 64 to km 74. In comparing this to the distributions of the bank materials (Figure 4.1) the significance of these trends becomes apparent. The foothills segment of the study reach contributes both locally derived material as well as those of Cordilleran origins. The reach from km 20 to km 34 has a number of bedrock outcrops which have at least partially deflected the course of the river. The same is true for the km 54 to km 75 sub-reach. It appears, therefore, that a significant amount of sediment production occurs where the bedrock outcrops. Second,

there is also an apparent lag effect whereby the sediment samples show a relatively high percentage of sandstone at considerable distances downstream from the source areas. This suggests that bedload transport in the river is of considerable importance. The fragile nature of the locally derived materials should limit the residence time of the larger pieces of sandstone within the river, before they are broken down into small fragments. These sandstone fragments would therefore not be expected to remain for more than a few seasons. Furthermore, in the foothills sub-reach the percentage of sandstone is relatively low. This may be explained by two possible factors. The first is that sediment transport activities may be of sufficient magnitude to rapidly reduce the size of local clasts. The second is that variations in bedrock lithology may produce a relatively high proportion of locally derived siltstone, clay ironstone, chert and conglomerate clasts. This would only account for a small portion of the residual percentage (Figure 4.11) though, and thus it appears probable that bedload transport and attrition may be comparatively efficient in the foothills sub-reach.

Figures 4.12 to 4.15 show scattergrams and regression lines of the mean B axis length and standard deviations plotted against distance downriver. Figure 4.12 A and B, represent the combined lithologies. The slope of the regression line is significantly steeper for the foothills sub-reach than for further downstream. On this basis the following discussion relates to the two sub-reaches so defined. The same trend is evident for the standard deviation, indicating that the sorting increases with distance downstream. Figures 4.13 to 4.15

Figure 4.12 A and B Scattergrams of mean size and standard deviation versus distance downstream for combined lithologies.


A
1 Y=59.5-0.44 X
2 Y=42.3-0.14X
SIZE (mm)
B
1 Y=42.5-0.42 X
2 Y=22.7-0.10 X
STANDARD DEVIATION (mm)
DISTANCE (KM)

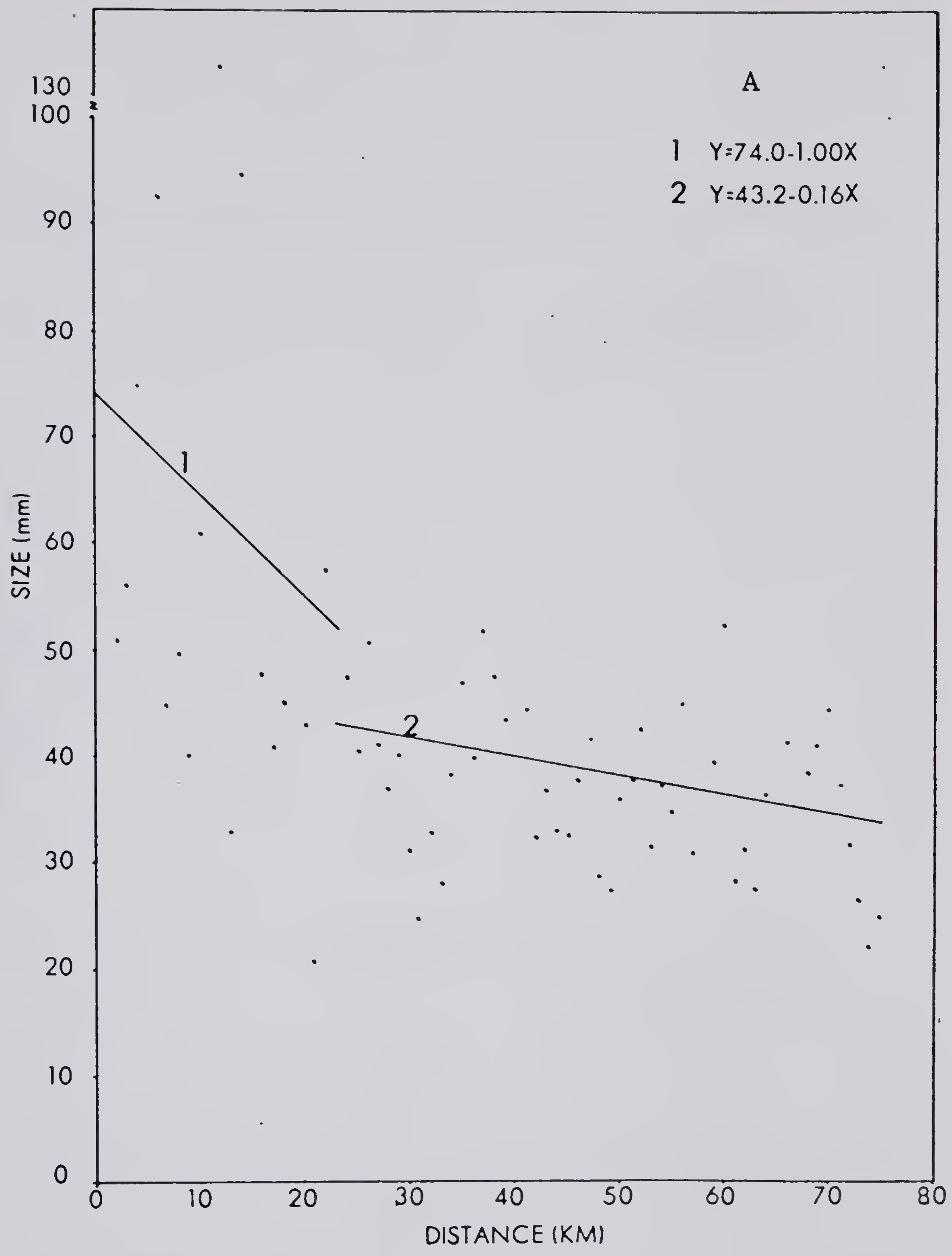


Figure 4.13 A Scattergram of mean size versus distance downstream for the quartzites.

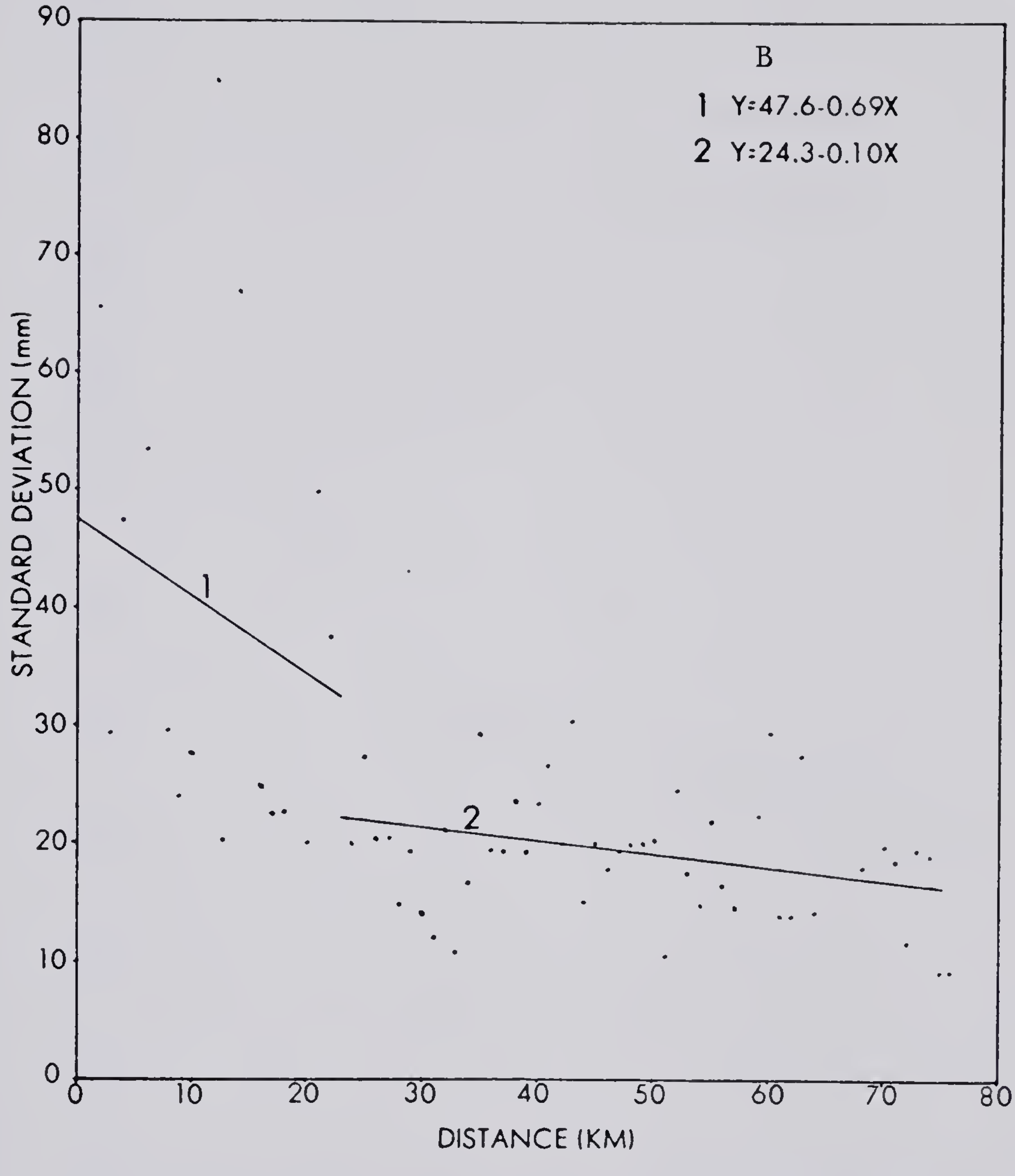


Figure 4.13 B Scattergram of standard deviation versus distance downstream for quartzites.

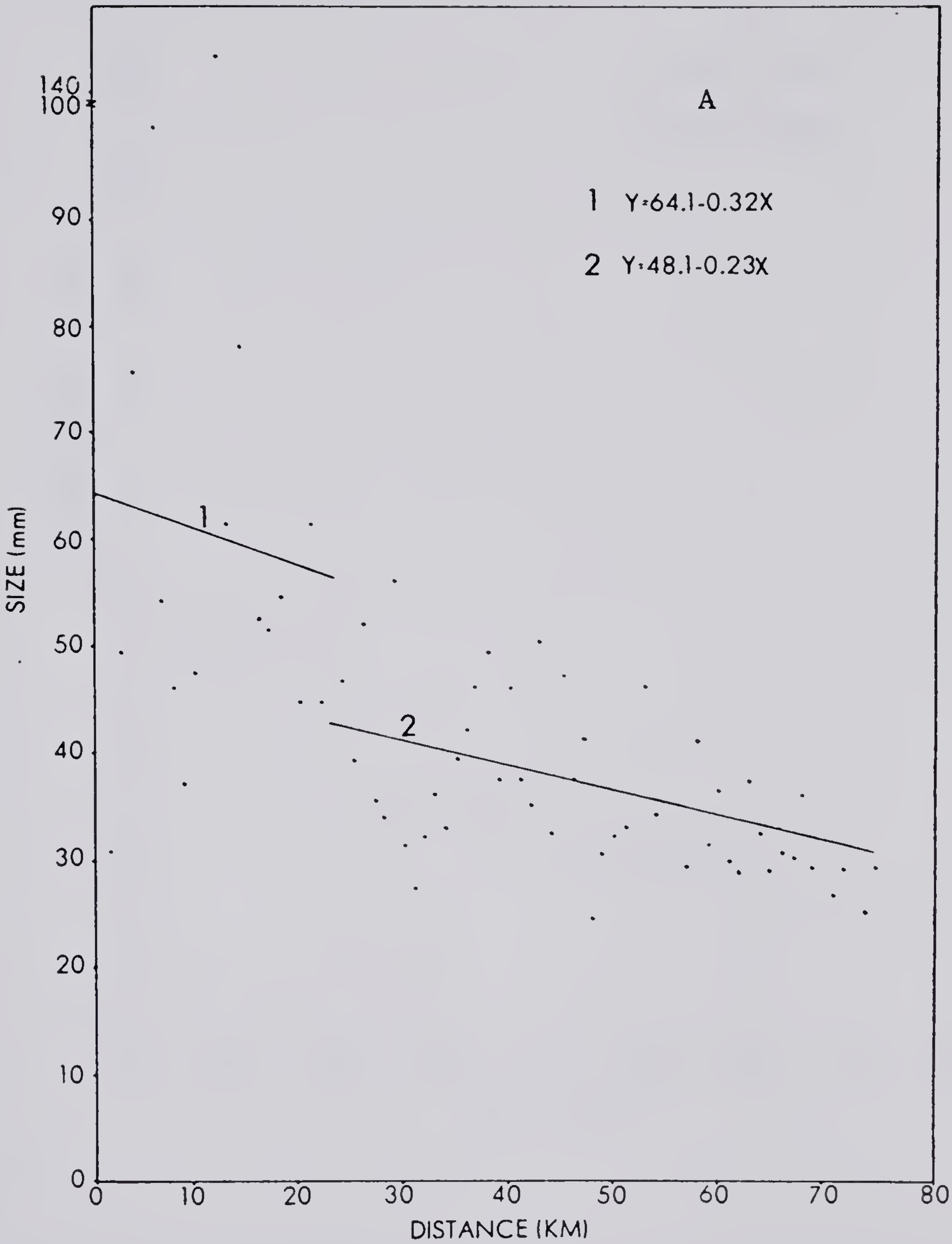


Figure 4.14 A Scattergram of mean size versus distance downstream for sandstones.

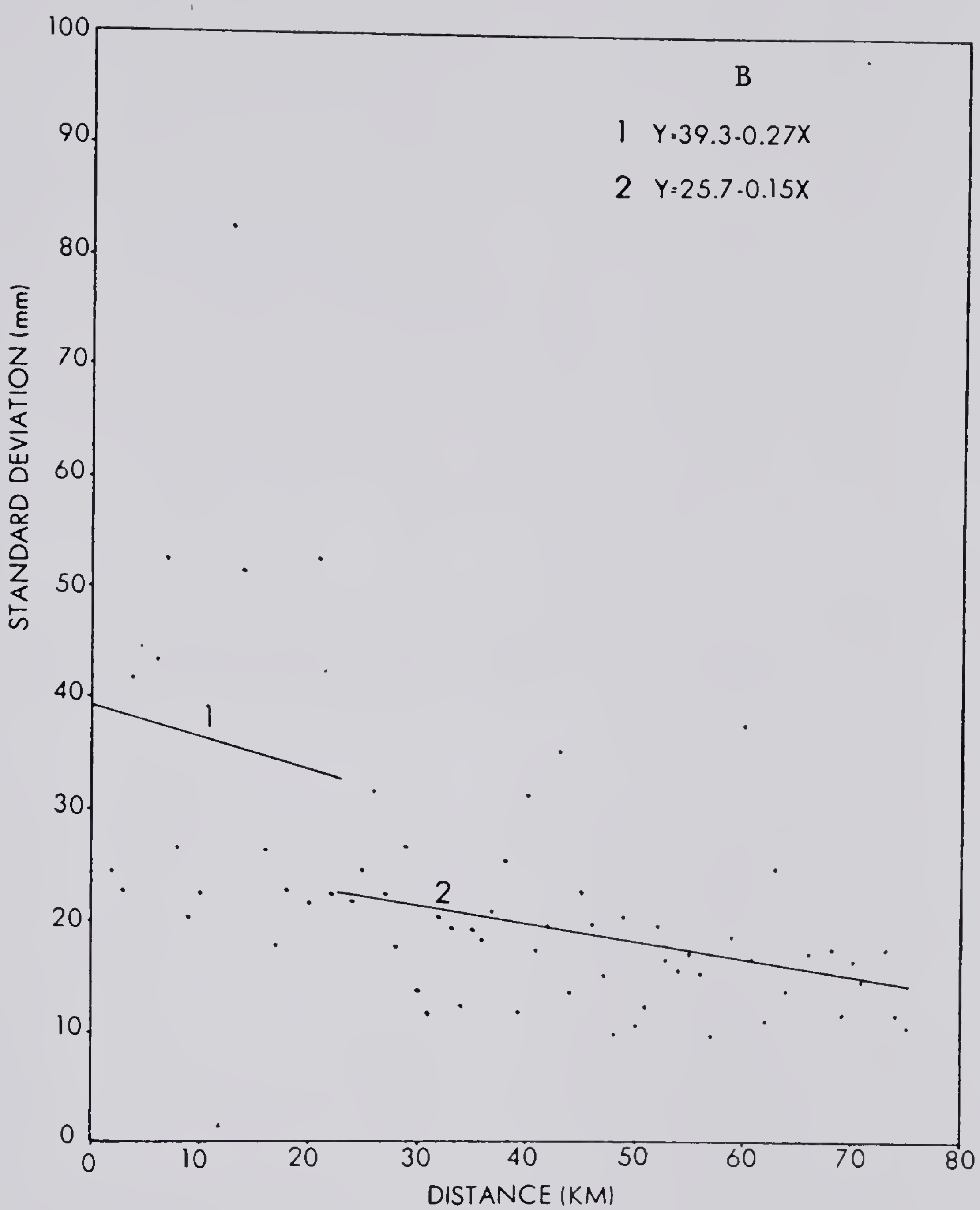
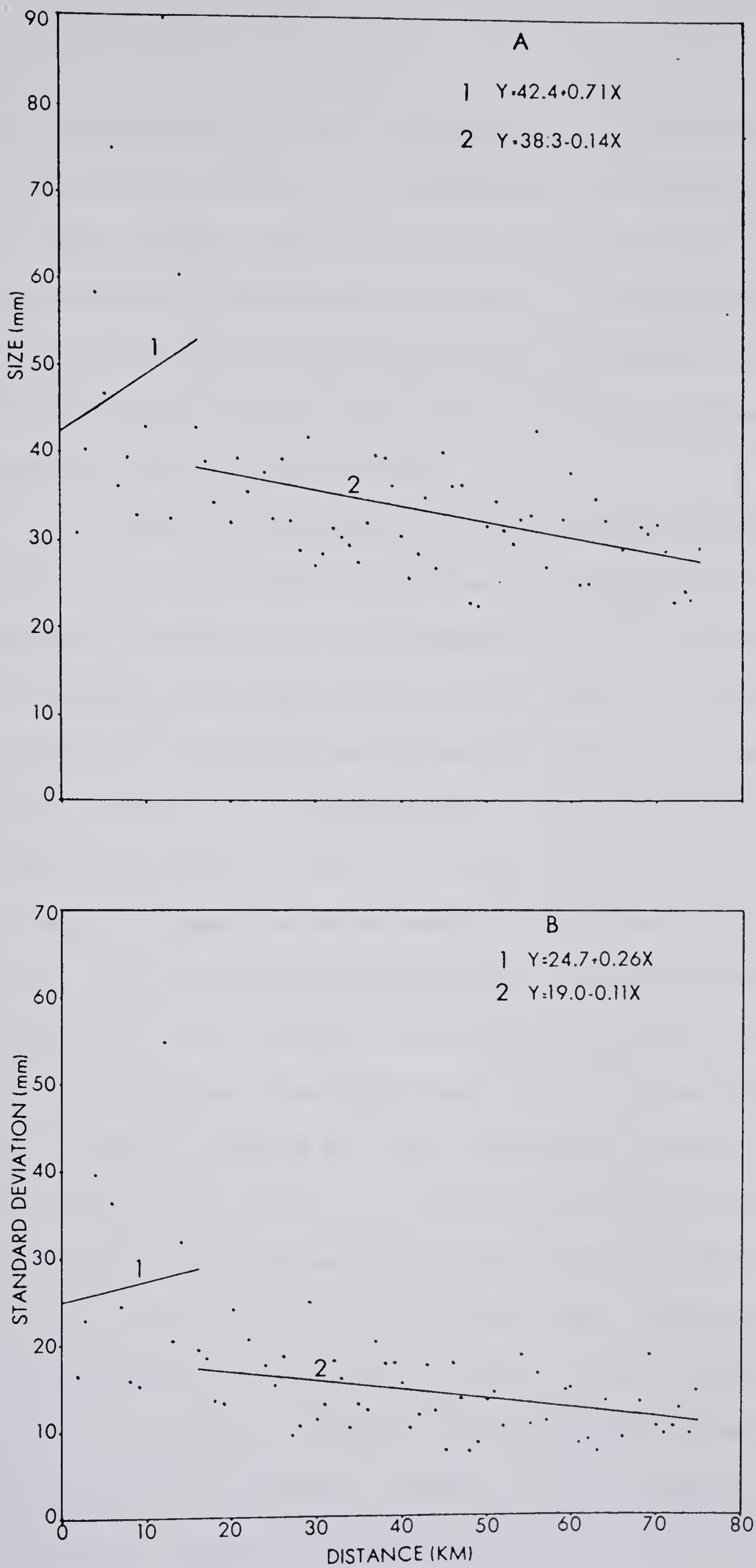


Figure 4.14 B Scattergram of standard deviation versus distance downstream for sandstones.

Figure 4.15 A and B Scattergrams of mean size and standard deviation versus distance downstream for carbonates.

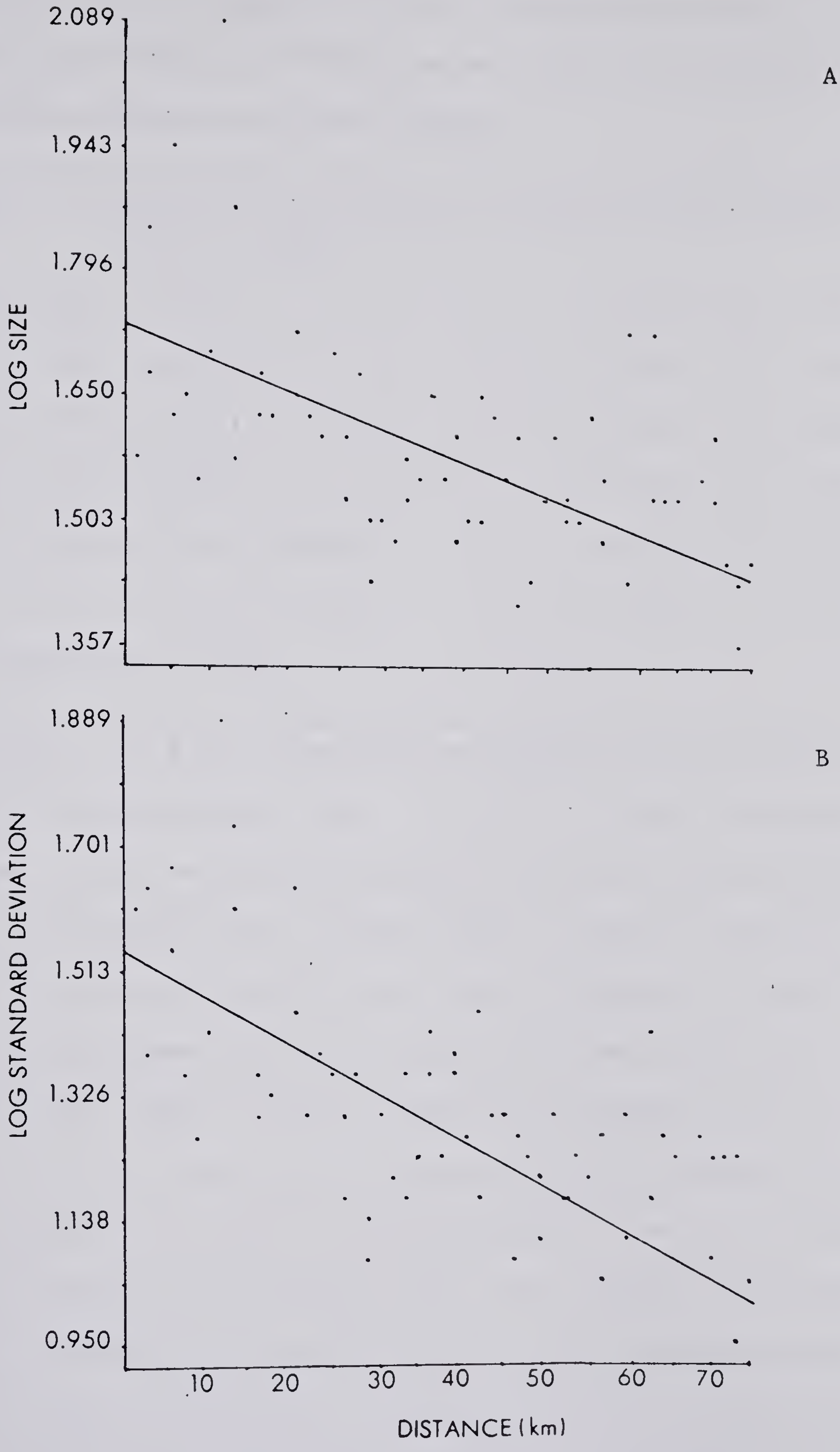

A
1 Y=42.4+0.71X
2 Y=38.3-0.14X
SIZE (mm)
B
1 Y=24.7+0.26X
2 Y=19.0-0.11X
STANDARD DEVIATION (mm)
DISTANCE (KM)

indicate the variations in size and sorting, of the three main lithologies, with distance downriver. The sandstones and quartzites show the same basic trends - that is, a higher rate of decrease occurs in the foothills than is experienced downstream. The carbonates, however, show that the mean size increases and the sorting decreases in the foothills segment of the study reach. No obvious explanation of this apparent anomaly is offered here.

The sorting of the material in the foothills sub-reach is generally poor, with large standard deviations occurring for all of the scattergrams. The quartzites and sandstones show the largest variations. The poor sorting of the sandstones may result partly from the addition of a large variety of clast sizes throughout this area. The quartzites and carbonates, to a lesser degree, also have a wide variety of clast sizes contributed to the river in this area.

Table 4.1 shows the relationships established for the combined gravel data of the two defined sub-reaches. The relationships which were obtained for the individual sub-reaches (as noted in Figure 4.12 to 4.15) were found not to be significant because of the small number of samples especially within the upper, foothills, section. It was decided, therefore, to apply a log transformation to the size and standard deviation data and perform a least squares regression analysis on all of the samples from the entire study reach. The degree of explanation obtained is very poor, with none of the r^2 values exceeding 50 percent. There is a tendency for the standard deviation to be better explained by the position and slope than is the mean particle size. A multiple regression analysis of mean size and standard devia-

Figure 4.16 A and B Scattergrams of log of mean size and log of standard deviation versus distance downstream for combined lithologies.


A
2.089
1.943
1.796
1.650
1.503
1.357
LOG SIZE
B
1.889
1.701
1.513
1.326
1.138
0.950
LOG STANDARD DEVIATION
10
20
30
40
50
60
70
DISTANCE (km)

tion against slope and position did not strengthen the degree of explanation significantly. It should be noted that the slope data used in the above analysis are not very precise.

Table 4.1 Results of the statistical analysis of the channel material samples.

Log Size = 1.72 − 0.003 X	r = 0.56	r^2 = 0.31
Log S.D. = 1.52 − 0.006 X	r = 0.66	r^2 = 0.44
Log Size = 1.41 + 40.90 S	r = 0.37	r^2 = 0.14
Log S.D. = 1.03 + 64.68 S	r = 0.41	r^2 = 0.17
Log Size = 1.66 + 10.94 S −0.003 X	r = 0.56	r^2 = 0.31
Log S.D. = 1.46 + 12.36 S − 0.005 X	r = 0.66	r^2 = 0.44

all significant to 99%

Figure 4.16 A and B show the scattergrams and calculated best fit lines for the transformed data. The scatter around the regression lines is still very high, especially for data of the foothills region. It is apparent that a larger number of samples would be required before firm conclusions could be drawn regarding the nature of the distributions, especially in this portion of the study reach.

The high degree of scatter around the regression lines of Figure 4.12 to 4.16, and the low r^2 values indicated in Table 4.1, may be explained by a number of possible factors. The first is that mentioned by Church and Kellerhals (1978) who suggested that large variations in gravel size and sorting occur within geomorphologically homogeneous areas. This possible source of variation is further compounded by the fact that the gravel samples were collected over an

[illegible]

extended period of time during which the discharge varied considerably. This may have caused some sampling bias. The second reason, and potentially the most important, relates to the source areas of the sediments. This problem was also disucssed by McPherson (1971) who related a high degree of data scatter to the addition of sediment from outside sources. In a situation such as that of the Upper Red Deer River study area, which has experienced a great variety of past geomorphic processes, the contribution of highly variable sediment sizes is assured. In addition to the variable sediments supplied to the river are the more subtle variations of local sedimentary environments at the sub-reach scale. Together these factors preclude a powerful explanation of the grain size and sorting distributions. The variations in the sorting of the material, recorded between the samples, may also be partly explained by the supplies of sediment being limited in the foothills, and by the probability that finer materials are transported out of the area leaving the coarser gravels as a pavement. In the lower reaches, where the river flows through the alluvial gravel deposits, channel migration across the floodplain ensures that a continuing supply of material is available for the development of channels, bars and islands.

The coefficient of diminution was derived for the combined type 1 gravel data using a method outlined by Shaw and Kellerhals (in press). This method requires a regression analysis of distance against the natural logarithm of the particle diameter. The slope coefficient, B, is then the coefficient of diminution (a_D). Comparisons of the two main sub-reaches within the study reach revealed

that the sub-reach from km 0 to km 23 has an a_D value of 0.0065 km^{-1} while the downstream section of the study reach has a value of 0.0039 km^{-1}. These two values compare favourably with a_D values found for other rivers given by Shaw and Kellerhals (in press). The values they established for some Albertan rivers include 0.0030 km^{-1} for the Red Deer River, 0.00182 km^{-1} for the North Saskatchewan River, and 0.00177 km^{-1} for the Bow-South Saskatchewan River system. The differences in the values calculated for the present study may be explained by differences in the average slope of the river (see Chapter Five) which decreases from 0.0051 in the foothills to 0.0037 downriver.

CHAPTER FIVE

CHANNEL, BAR AND ISLAND MORPHOLOGIES: CHANNEL MIGRATION

5.1 CHANNEL PATTERN CLASSIFICATION

5.1.1 INTRODUCTION AND METHODOLOGY

The problem of describing channel patterns in a meaningful fashion, relative to the controlling factors, has been recognised by several writers (see for example, Leopold and Wolman, 1957; Brice, 1964). A second problem is that rigid distinctions drawn between the three major channel pattern types; straight, meandering (sinuous), and braided, markedly obscure the reality of a channel pattern continuum. In response to this dilemma further subdivisions have been developed, such as that of Bray (1971). Although such subdivisions are slightly more satisfactory than the traditional three-fold classification, they are still rather subjective in their methodology and do not embrace the complete spectrum of natural channel patterns. For example, Leopold and Wolman (1957) determined channel patterns through the use of a sinuosity index,

$$S.I. = \frac{L_T}{L_V} \qquad (2)$$

where L_T is the length of the thalweg and L_V is the valley length,

[illegible]

[illegible]

[illegible]

[illegible]

[illegible]

[illegible]

of a particular reach. The index, therefore, differentiates between straight and sinuous channels. This was illustrated by Schumm (1977, p. 113 and 114) who commented that any such distinctions between the two patterns is purely arbitrary. The S.I. may only be applied with difficulty to braided channels because the problem of stage dependency arises. The selection of a reference stage, such as at bankfull discharge, for a braided reach is also unrealistic because of the merging of many channels at relatively high flow stages.

A more appropriate method was formulated by Brice (1964) whose classification centered on the differentiation of braided sections using the total length of the bars and islands within the reach. This braiding index is defined as;

$$B.I. = 2 \frac{\sum L_B}{L_R} \qquad (3)$$

where $\sum L_B$ is the total length of the bars and islands and L_R is the length of the reach. Brice (1964) conceded that the index has little hydraulic significance, especially with respect to channel(s) width, but that it does serve as a consistent discriminator of braided and nonbraided reaches. This method, also, has a number of deficiencies. The first is that it, too, is stage dependent and comparisons of separate braided reaches using aerial photographs must be based on a knowledge of the local discharge/stage relationships at the times of photography. This problem may be partly overcome by using the Stabilised Braiding Index (S.B.I.), suggested by Brice (1964), which considers only the stabilised islands in a particular reach. A second problem

is that, although the B.I. may adequately differentiate braided and nonbraided reaches, it does not distinguish major variations of single channel patterns.

In an attempt to partly overcome these problems a simple scheme was devised, for the present study, based on the width of the active channel. The basic assumption made here is that the width, at-a-station, will vary with the channel pattern, so that a straight single-channel reach will be narrower than a meandering one, all other variables remaining constant. This index will be referred to as the width ratio (Wr) and defined as;

$$Wr = \frac{W_S}{W_G} \qquad (4)$$

where W_S is the width, at-a-station, and W_G is the greatest at-a-station width in the study reach. The reference width used in this study is for bankfull stage at km 73. This was distinguished by marked discontinuities of the vegetation patterns. The surfaces below the bankfull stage heights have a sparse vegetation cover consisting of grasses and scattered willows. The surfaces above this stage limit have a more complex and better developed vegetation assemblage dominated by aspen poplar and spruce trees.

5.1.2 STUDY REACH CHANNEL PATTERNS

Figure 5.1 shows the average Wr values for each kilometre of valley length. On the basis of varied channel patterns the study reach may be subdivided into five major zones. The first, Zone 1,

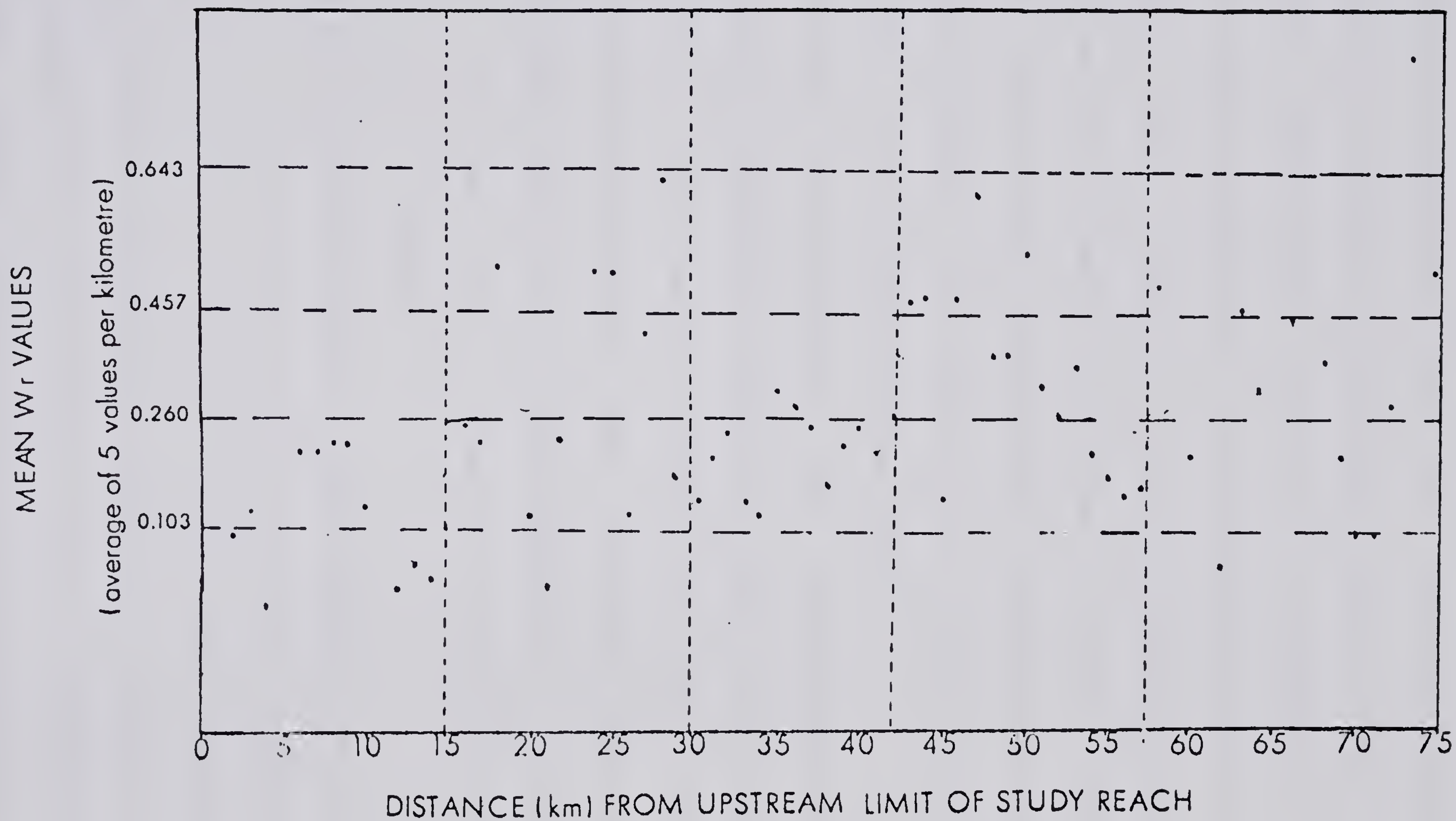


Figure 5.1 Distribution of Wr values with distance from the upstream limit of the field area.

is characterised by low Wr values with a gradual increase downriver. This zone is contained within the foothills area. The second division, Zone 2, is located between km 15 and km 29. This sub-reach experiences a large variation in the Wr values but again shows a trend of increasing widths downriver. Zone 3, from km 29 to km 42, indicates a moderating trend, with a slight increase in the width downstream and a relatively small amount of scatter of the Wr values. The fourth zone, located between km 42 and km 58, reveals a decrease in the active channel width through this sub-reach, although there is also a considerable amount of width variation. The final zone, 5, shows a highly variable width with trend of increasing width down-river.

The channel in the foothills, Zone 1, is incised into bedrock except for the area between km 6 and km 10, which is mainly bordered by gravels. Zones 2 and 3 border the Bearberry Prairie and are predominantly contained within gravel banks. The fourth zone, maintains predominantly gravel banks but is also influenced by a series of bedrock outcrops. The final sub-reach, Zone 5, is flanked by gravel with some minor outcrops of bedrock.

In more traditional terms the zones demarcated in this manner may be classified into pattern divisions of braided, meandering and straight. Zone 1 embodies a primarily straight single-channel, while that of Zone 2 is highly variable but dominated by braided sub-reaches. The third zone maintains a single-channel pattern of varying sinuosities. The fourth zone is braided, although the braiding appears to decrease in intensity downstream. From km 54 to km 56 the river has a single-channel pattern. In Zone 5 braiding dominates but, as in Zone 2,

there are sub-reaches which are braided while other sub-reaches are of single-channel form.

Several smaller sub-zones may be delineated. For example, in Zone 1 an area of higher Wr values occurs between km 6 and km 10. In Zone 3 a division occurs at km 38. The sub-section upstream from this point is variable in width, although the variations are not as great as in the previous section. Downriver the width is more uniform.

A final generalisation relates to the apparent association between the Wr values and the occurrences of bedrock outcrops along the floodplain margins. In the areas where bedrock outcrops the river widths are relatively constricted. Directly upriver from these points the widths tend to be much greater, with sharp transitions between the narrow and wide sub-reaches. This theme will be discussed further in Chapter Six.

5.2 BAR AND ISLAND CLASSIFICATION

5.2.1 INTRODUCTION AND METHODOLOGY

Several attempts have been made to produce classifications of landforms in river channels, especially active bar types. These classifications are based mainly on criteria such as bar shape, relative location in the channel (with respect to the banks) and the orientation to the direction of flow. A second type of classification is based on the dominant mode of formation. This type of classification distinguishes between those forms of mainly depositional origin and

those which have a more complex depositional-erosional history (Smith, 1974). The latter approach was not considered appropriate for the present study because of the apparently complex nature of bar formation, involving simultaneous deposition and erosion of fluvial sediments, in the upper Red Deer River.

Table 5.1 summarises numerous published classifications of alluvial bar types. The first seven bar types listed are essentially mid-channel forms, separated from the main banks. The activity of each of these forms depends largely on the hydraulic regime of the river. The differentiation of these bar types is not always consistent, sometimes being based on sediment size relationships, in other cases on shape and orientation relative to flow directions. The criteria of bar shapes and orientation to the direction of flow are of somewhat dubious value because both are strongly stage dependent. A high stage flow will often leave only a small area of a bar, if any, exposed while the same bar at a lower stage may fit a different classification. For example, field observations of the upper Red Deer River showed that diagonal bars at high flows might be classified as small erosional remnants on point bar surfaces at lower flow stages. Because all alluvial bars have diffuse boundaries and irregular surfaces their categorisation may never be precise.

Types 8, 9 and 10 (Table 5.1), tributary, point and side bars, are convergent with channel banks. Side bars are usually associated with straight or slightly sinuous channels (Kellerhals, et.al., 1976; Schumm, 1977) while point bars are particularly characteristic of

Table 5.1 Bar classification based on published descriptions

Bar Type	Location	Shape	Relative Stability	Sources
Longitudinal bar	Midchannel, straight shallow reaches	Elongate-convex upward surface	Active	Smith (1974)
Transverse bar	Midchannel, isolated, occupy nearly full width	Straight, lobate, broad flat surface	Active	A.S.C.E. Task Force (1966); Allen (1968); Smith (1974).
Diagonal bar	Midchannel, straight reaches	Varied	Active	Church (1972); Kellerhals et al., (1976).
Spool, Braid Diamond bar	Midchannel straight reaches	Spool or delta-shaped	Active permanent (if vegetated)	Krigstrom (1962); Allen (1968); Church (1972); Kellerhals et al., (1976).
Midchannel bar	Midchannel	In lee of obstructions	Active (may become stabilized)	Kellerhals et al., (1976).
Overflow bar	a) Straight reaches, mid channel b) Extending downstream from outside bank	Horse-shoe shaped	Active	Krigstrom (1962).
Linguoid bars	Occur anywhere in the channel, staggered arrangement	Slio-off slope in downstream direction, lobate to wedge shaped	Active	Allen (1968); Collinson (1970); Kellerhals et al., (1976).

Table 5.1 (continued)

Bar Type	Location	Shape	Relative Stability	Sources
Tributary bars, Channel Junction bars	Junction of two channels or at the mouth of a tributary out into channel forcing water towards the outside bank	Initial stages-low flat, tongue shaped with growth-extends out into channel forcing water towards the outside bank.	Active	Krigstrom (1962); A.S.C.E. Task Force (1966); Kellerhals et al., (1976).
Side bars, alternating bars	Attached to alternate banks, in straight reaches	Gradual widening downstream. Lateral extent less than width	Active	A.S.C.E. Task Force (1966); Allen (1968); Collinson (1970); Kellerhals et al., (1976).
Point bars	Attached to inside of meander bend	Lateral extent less than width	Permanent	Krigstrom (1962); A.S.C.E. Task Force (1966); Allen (1968); Kellerhals et al., (1976).
Islands	Midchannel	No fixed shape	Permanent vegetated	Kellerhals et al., (1976).

well developed meanders (Kellerhals, et al., 1976). Tributary bars are located downstream from the confluence of a tributary.

Islands (Type 11, Table 5.1) are relatively permanent features with varying degrees of vegetation established on their surfaces. They occur in midchannel and are sometimes found in actively braiding rivers in association with many of the bar types mentioned earlier.

The numerous bar types listed in Table 5.1 involve a cumbersome and imprecise terminology. Misinterpretation of certain forms is quite possible from the varying descriptions outlined in the related literature. A major problem is that nowhere is a reference flow stage mentioned. Bar forms identified for one flow stage may take on quite different shapes and relative orientations at other stages. The strict application of a reference stage is not always possible in studies of river morphologies. If aerial photographs are used, particularly sequential photographs, for times of different discharges an accurate assessment of bar evolution is rendered difficult, if not impossible.

For this study a simplified classification is employed. It distinguishes, at the most general level, midchannel bars from those attached to the channel banks and includes an island category. Further subdivisions of the main classes are based on a four-fold bar type scheme, two types of islands, and the relative stability of these. A qualitative assessment of their relative stability was based on the degree to which they were vegetated because no direct measurements of migration activity could be made.

Table 5.2 Bar and island classification used in this study

	Level 1 Class	Level 2 Bar Type	Level 3 Relative Stability
		Longitudinal Bar	Very Stable
1.	Mid Channel		Stable
		Braid Bar	Unstable
		Side Bar	Very Stable
2.	Channel margin (= Attached)		Stable
		Point Bar	Unstable
		Island Type	
		High surface	
3.	Islands	relative to active floodplain. (at or above bankfull stage)	Very Stable
		Low surface	
		relative to active floodplain. (below bankful stage)	Stable

River islands have been rather neglected in the geomorphological literature. These forms were defined by Kellerhals, et.al., (1976) as being relatively stable and frequently vegetated. They may be formed by the coalescence and stabilisation of a number of bars, or by successive deposition on a bar, until much of its surface is almost permanently above the river level. The latter may be accomplished by extreme floods of moderate to long recurrence intervals depositing alluvium on the island surface. Another mode of island formation, common in the upper Red Deer River Valley, is the isolation of large segments of the floodplain by the reactivation of former braid channels.

5.2.2 STUDY REACH BARS AND ISLANDS: CHANNEL MIGRATION

Based on the criteria of variations in bar and island form, and channel migration, the study reach may be divided into a number of major sub-units. The bar and island descriptions are based on field observations, while the channel migration evidence is derived from interpretation of aerial photographs taken between 1957 and 1972. The cross sections were originally surveyed in 1976 by Water Surveys, Alberta Environment, and partly resurveyed during the 1978 field season. The 1978 surveys were conducted at seven different sites, which yielded a total of seventeen channel cross sections. The sites were located downstream from the foothills zone and were chosen to adequately represent a variety of channel patterns. Where possible three sections were surveyed at each site. At some locations this was not possible

because of very dense vegetation cover. In these situations cutlines were utilised and one section surveyed. The surveying was accomplished using a jet boat and crew supplied by Water Surveys, Alberta Environment and a recording depth sounder. The exposed bars and islands were incorporated into the cross section surveys by use of a level and stadia rod. Unfortunately the absolute heights of these sections could not be determined. Each cross section is thus referenced to an estimated, local datum.

a) SUB-UNIT 1 (km 0-km 15), FIGURE 5.2 A TO H

This zone extends the length of the foothills sub-reach. The landforms found in this sub-reach include islands, midchannel, longitudinal and side bars. The islands were classified as the high surface type, located at and above the bankfull level and vegetated with species similar to those found on the floodplain-spruce and aspen poplar (Figure 5.2 C and D). These islands are composed of gravel with a thin (less than 0.5 metre thick) layer of finer material. The bars found along this reach are numerous and vary in size. They are composed of gravel and located well below the bankfull level (Figure 5.2 E). Figure 5.2 F to H show three channel cross sections (Cross Sections 1, 2 and 3) from this sub-reach. Cross Section 1, located at km 4, indicates a multiple-channel reach with a number of the high surface islands and side bars. Cross Sections 2 and 3 show a single-channel reach entrenched into bedrock. Attached bars predominate in this section of the unit and clearly defined limits of the bankfull stage are not recognisable.

Channel migration in this sub-unit does not appear to have

Figure 5.2 A and B Floodplain modification in sub-unit 1.

A is an interpretation based on aerial photographs from 1958 (scale approximately 1:16,000).

B is an interpretation based on aerial photographs from 1972 (scale approximately 1:21,000).

Scale of A and B approximately 1:3,000

LEGEND

Exposed gravel

Perennial grasses with scattered stands of willow and immature aspen poplar

Dense stands of aspen poplar

Stands of spruce and some aspen poplar

The first class encompasses the major bar forms mentioned in the text. The second group includes the low surface islands with a transition to the high surface islands which are mainly contained within the last two classes.

Discharge at the time of photography (from the gauging station near Sundre): A = 3,000 c.f.s. (85 c.m.s.)

B = 591 c.f.s. (16.7 c.m.s.)

1 —— Channel Cross-Section (note cross-sections two and three not located)

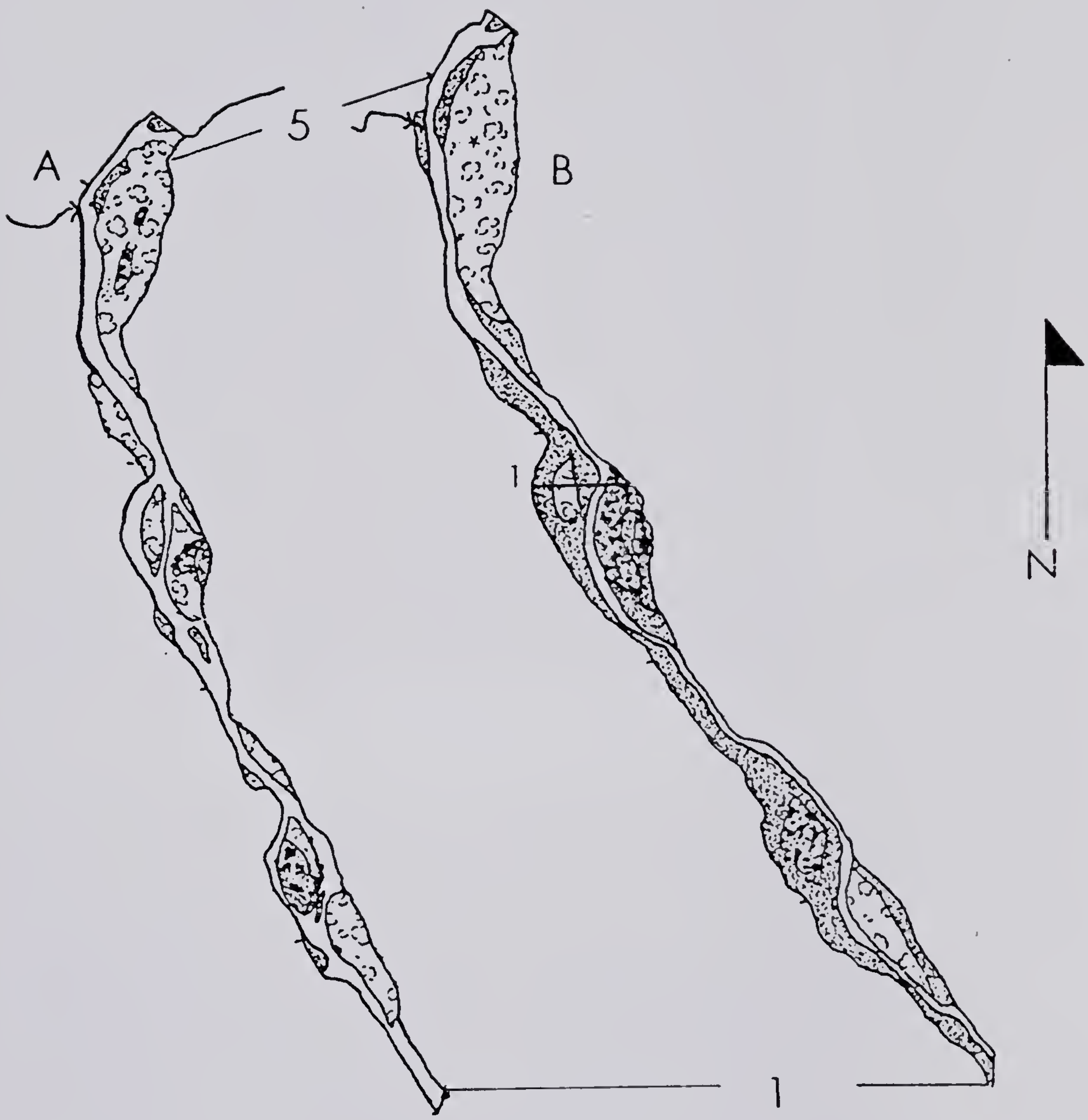

A
5
B
1
1
N

Figure 5.2 C and D Oblique, aerial photographs of the upper Red Deer River in the foothills region. Note the well vegetated islands (1) and extensive exposed gravel surfaces.

C

D

Figure 5.2 E A small midchannel longitudinal bar located in sub-unit 1.

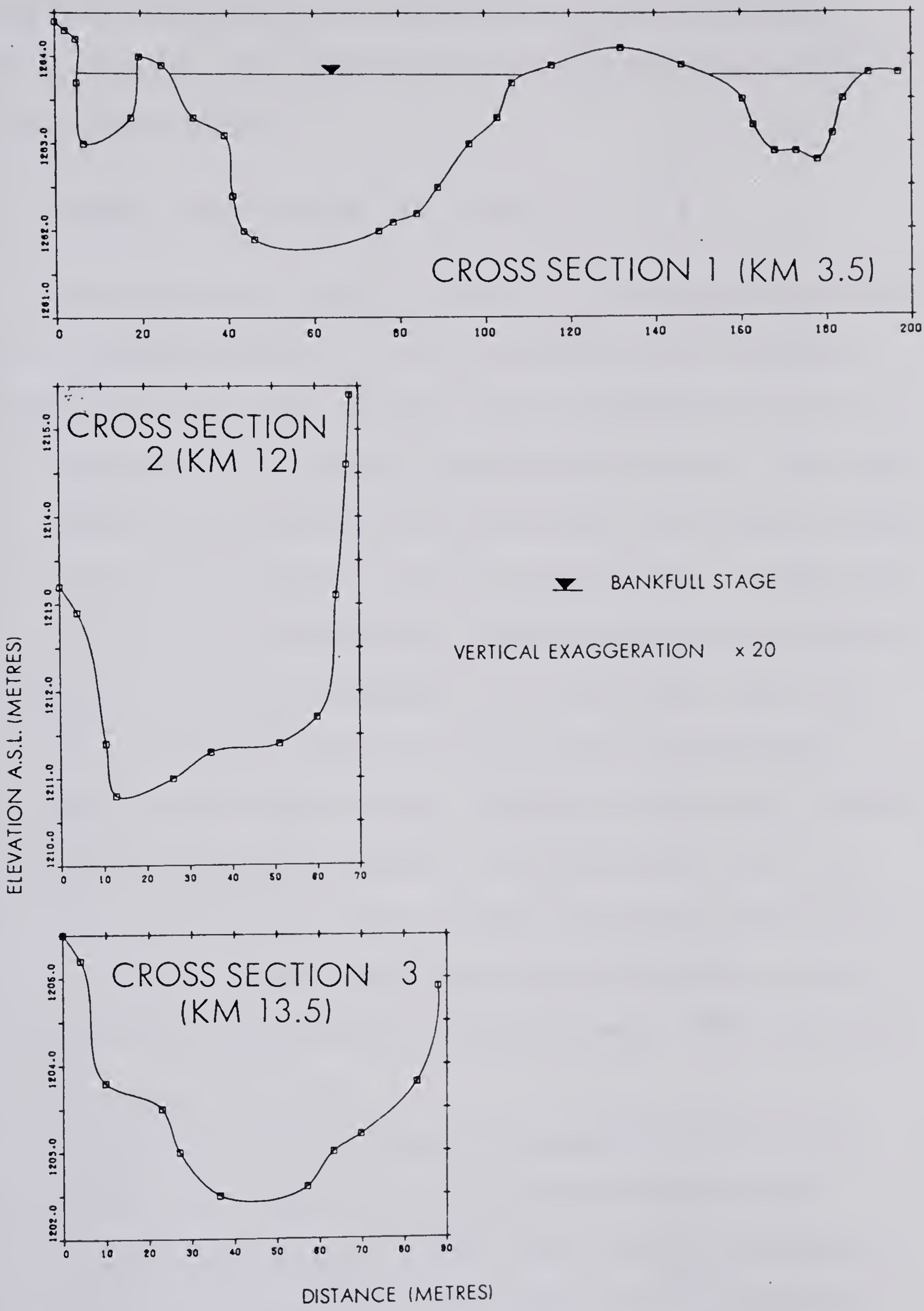


Figure 5.2 F,G and H Cross sections 1,2 and 3.

been extensive during the years separating the aerial photographs (Figure 5.2 A and B). The islands have not changed in size or shape to any significant extent.

b) SUB-UNIT 2 (km 15 TO km 32), FIGURE 5.3 A TO K

As the river moves from the foothills onto the Bearberry Prairie a number of changes take place in the morphology of the floodplain. The islands appearing in this sub-reach are of both the high and low surface categories and are composed predominantly of gravel. The high surface islands have a thin layer of finer material overlying the gravel but this layer does not appear on the low surface forms. In this sub-unit, as in the first, the high surface islands are vegetated with spruce and aspen poplar, as is the floodplain. The low surface forms have a less developed vegetative cover, consisting mainly of willow and immature aspen poplar (Figure 5.3 C). Examples of high surface islands are illustrated by Figure 5.3 D and E. A concentration of the low surface forms occurs in the central portions of the reach from km 23 to km 34. The high surface forms are distributed throughout but are most abundant in the upper sections of the sub-reach. Their locations are shown in Figure 5.3 A and B.

Numerous bars are also evident, encompassing the four types outlined in Table 5.2. These bars are composed of gravel and are sparsely vegetated with perennial grasses and a few poorly developed, scattered stands of willow (Figure 5.3 F and G). The point bars are not the "classical" forms described in the literature. They are seasonally active gravel deposits located on the insides of bends in the river.

They have been dissected in a number of cases by seasonally active, secondary channels. The vegetative cover of the point bars is generally sparse, with some small stands of willow and aspen poplar usually occurring close to the floodplain-bar boundary.

Figure 5.3 H to K show selected channel cross sections from this sub-reach (Cross Sections 4, 5, 6 and 7). Cross Section 4 is located in a straight reach bordered on one side by bedrock. An attached side bar is found in this section separated from the bank by a small secondary channel. Cross Sections 5 and 7 show the various point bar shapes. Section 5 is located near km 22 where a bar surface has been incised by secondary channels. A small bench-like form, possibly of erosional origins, is located on the surface. Cross Section 7 shows a well-developed point bar-island complex which is at present being incised and has been separated from the main banks by two permanent channels. The shape of this form suggests a more stabilised surface than the bar previously described, as it is lacking secondary channels. The area shown at Cross Section 6 is located at a complex of both high and low surface islands.

The channel migration activity in this unit has been relatively high. Numerous gravel surfaces have been reactivated and incised to form low surface islands (Figure 5.3 A, B, J and K). This reactivation has resulted from major channel migration and displacement of the main channel, evident throughout the entire sub-reach. An area where active point bar formation has occurred in the period of time separating the aerial photographs is in the vicinity of km 22.

Figure 5.3 A and B Floodplain modification in sub-unit 2

A is an interpretation based on aerial photographs from 1963 (scale approximately 1:30,000).

B is an interpretation based on aerial photographs from 1972 (scale approximately 1:21,000).

Scale of A and B is approximately 1:3,000.

LEGEND

Exposed gravel

Perennial grasses with scattered stands of willow and immature aspen poplar

Dense stands of aspen poplar

Stands of spruce and some aspen poplar

The first class encompasses the major bar forms mentioned in the text. The second group includes the low surface islands with a transition to the high surface islands which are contained within the last two classes.

Discharge at the time of photography (from the gauging station near Sundre): A = 437 c.f.s. (12.4 c.m.s.)

B = 591 c.f.s. (16.7 c.m.s.)

5 ——— Channel Cross-Section

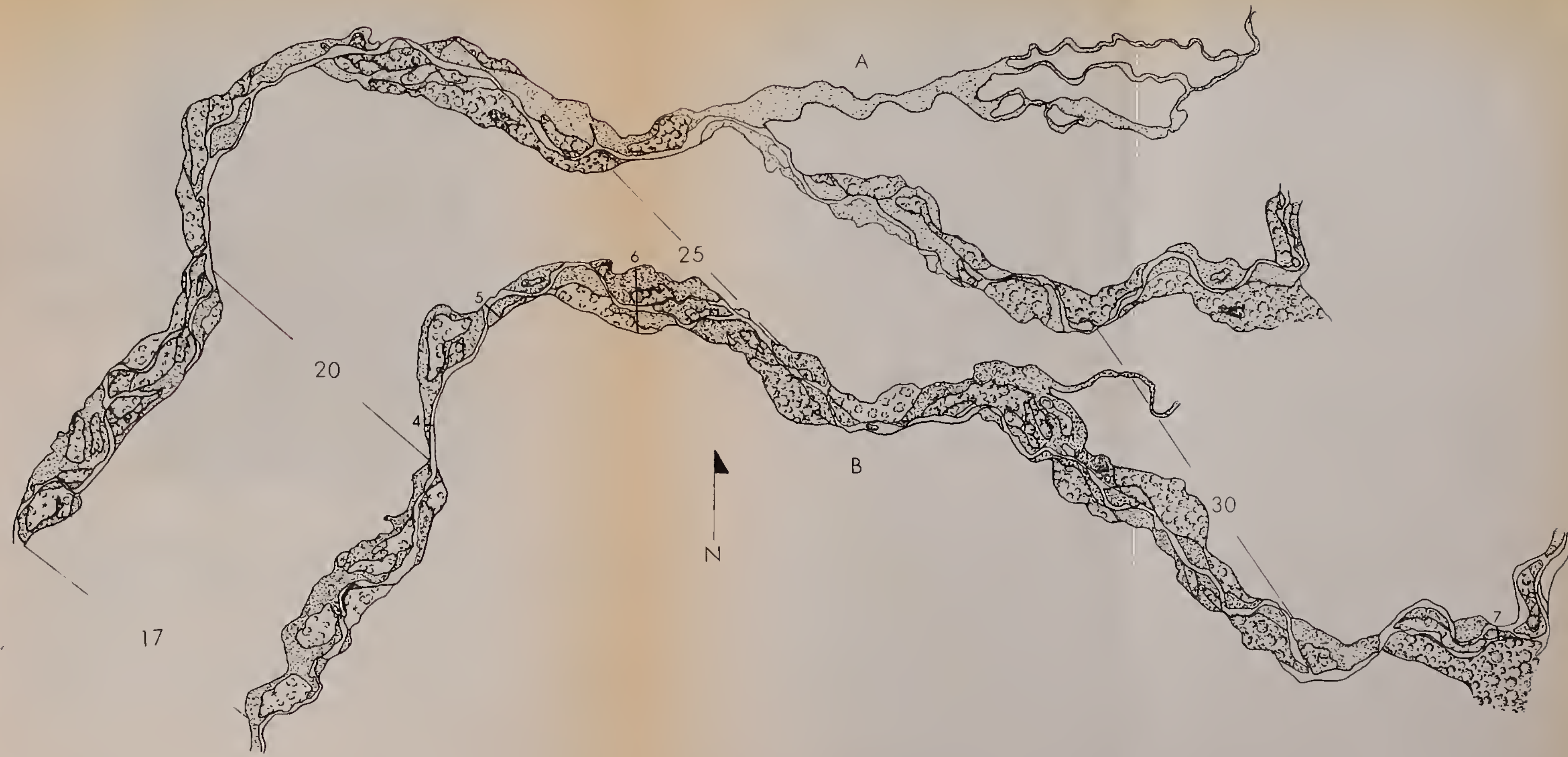

A
B
N
25
30
20
17
6
5
4
7

Figure 5.3 C Oblique, aerial photograph of a low surface island. Note poorly developed vegetation on the surface.

D

E

Figure 5.3 D and E High surface islands (1) contained in sub-unit 2. Note large exposed gravel surfaces (2) separating these islands from the main banks.

Figure 5.3 F Low-angle, oblique, aerial photograph of a bar and island complex. The island is a stabilized and modified braid bar.

Figure 5.3 G An exposed gravel surface on the upstream edge of an island.

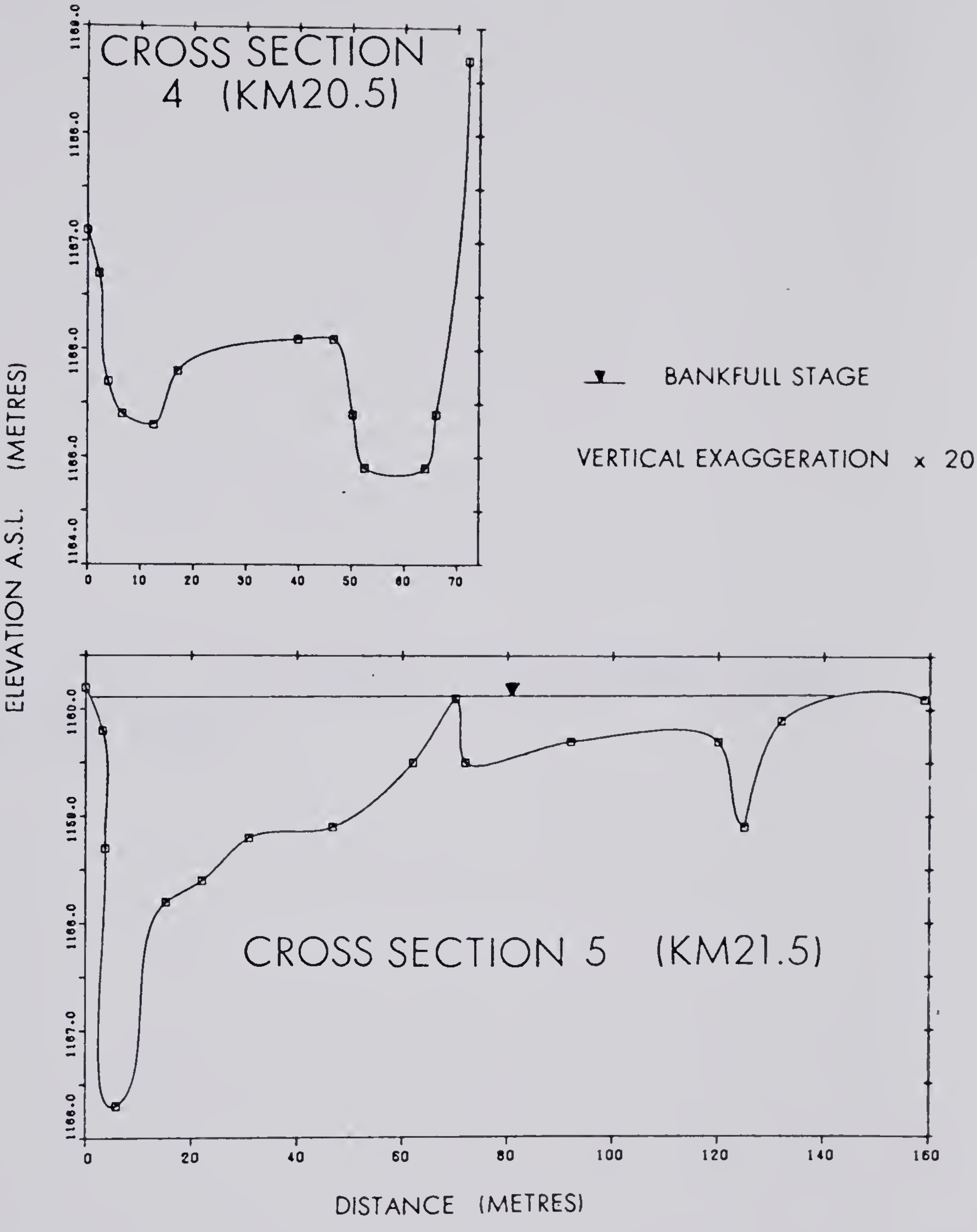


Figure 5.3 H and I Cross sections 4 and 5.

Figure 5.3 J Cross section 6 (note change of scale).

(see back jacket)

Figure 5.3 K Cross section 7 (note change of scale).

(see back jacket)

c) SUB-UNIT 3 (km 32 TO km 39), FIGURE 5.4 A to E

This short reach is located upstream from Sundre. Islands are scarce but those which do occur are of the low surface type. They are composed predominantly of gravel, without a surface layer of finer material. The vegetative cover consists mainly of perennial grasses and scattered stands of willow and aspen poplar (Figure 5.4 D).

The bars formed in this sub-reach are predominantly of the attached point bar variety with a few midchannel longitudinal bars (Figure 5.4 C). The variety of point bars found in this reach have also been observed by several previous workers on rivers in Great Britain and Texas (Lewin, 1976; Gustavson, 1978). At their present stage of development they may be considered as embryo point bars (Bluck, 1976). Figure 5.4 E (Cross Section 8) shows a channel cross section located across one of these forms. If allowed to develop fully, these bars will be similar to those described for the previous sub-reach.

The sub-reach has not experienced pronounced channel shifting during the years separating the aerial photographs (Figure 5.4 A and B). A few isolated examples of point bar dissection were recognised and these are concentrated in the lower portions of the sub-reach. The rate of bar formation, notably point bar formation, appears to have been lower than in the previous sub-reach.

d) SUB-UNIT 4 (km 39 TO km 42), FIGURE 5.5 A to I.

The islands in this sub-reach may be divided into both high and low surface varieties. Those of the high surface type have sedimen-

Figure 5.4 Floodplain modification in sub-unit 3.

A is an interpretation based on aerial photographs from 1963 (scale approximately 1:30,000).

B is an interpretation based on aerial photographs from 1972 (scale approximately 1:21,000).

Scale of A and B is approximately 1:3,000.

LEGEND

Exposed gravel

Perennial grasses with scattered stands of willow and immature aspen poplar

Dense stands of aspen poplar

Stands of spruce and some aspen poplar

The first class encompasses the major bar forms mentioned in the text. The second group includes the low surface islands with a transition to the high surface islands which are contained within the last two classes.

Discharge at the time of photography (from the gauging station near Sundre): A = 437 c.f.s. (12.4 c.m.s.)

B = 591 c.f.s. (16.7 c.m.s.)

8 —— Channel Cross- Section

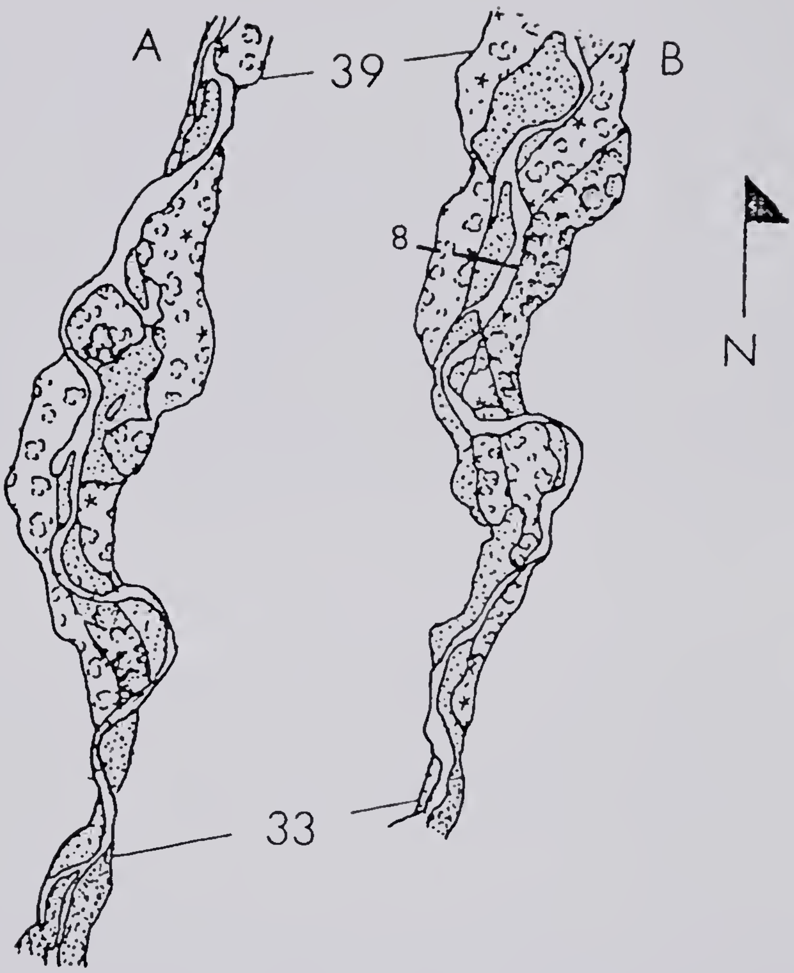

A
39
B
8
N
33

Figure 5.4 C Oblique, aerial photograph showing a sinuous portion of the study reach. Flow is away from the observer.

Figure 5.4 D A dissected point bar surface at km. 35. Looking west towards Bearberry Prairie.

Figure 5.4 E Cross section 8

tary and vegetative characteristics similar to the floodplain (Figure 5.5 E). The low surface islands are composed of gravel and are vegetated with perennial grasses (Figure 5.5 F).

Some well developed point bars are found in this sub-reach (Figure 5.5 C) and are concentrated in its upstream portion. A number of longitudinal bars are associated with them. Downstream, there is an increase in the number of midchannel spool bars and islands (Figure 5.5 D). Figure 5.5 and 1 (Cross Sections 9 and 10) show two channel cross sections within the sub-reach. Cross Section 9 is located at a dissected point bar and Cross Section 10 is located further downstream where the braiding is more intense. Note the occurrence of several high surface islands located at and above the bankfull level.

Channel migration activity in this sub-reach has been generally much higher than in the previous sub-reach (Figure 5.5 A and B). Furthermore, a higher rate of point bar formation appears to have occurred in this sub-reach compared to the previous one. In the lower sections of the sub-unit point bar dissection and the development of low surface islands dominate. In the upper portion of the sub-reach the development of a few high surface islands has taken place. This has also been accompanied by the development of a relatively large, active, cross sectional width.

e) SUB-UNIT 5 (km 42 TO km 75), FIGURE 5.6 A TO L

This long sub-reach has a comparatively full spectrum of bars and islands. Islands of both high and low surface varieties are found throughout and are not concentrated in one area. The high surface

Figure 5.5 Floodplain modification in sub-unit 4.

A is an interpretation based on aerial photographs from 1963 (scale approximately 1:30,000).

B is an interpretation based on aerial photographs from 1972 (scale approximately 1:21,000).

Scale of A and B is approximately 1:3,000.

LEGEND

Exposed gravel

Perennial grasses with scattered stands of willow and immature aspen poplar

Dense stands of aspen poplar

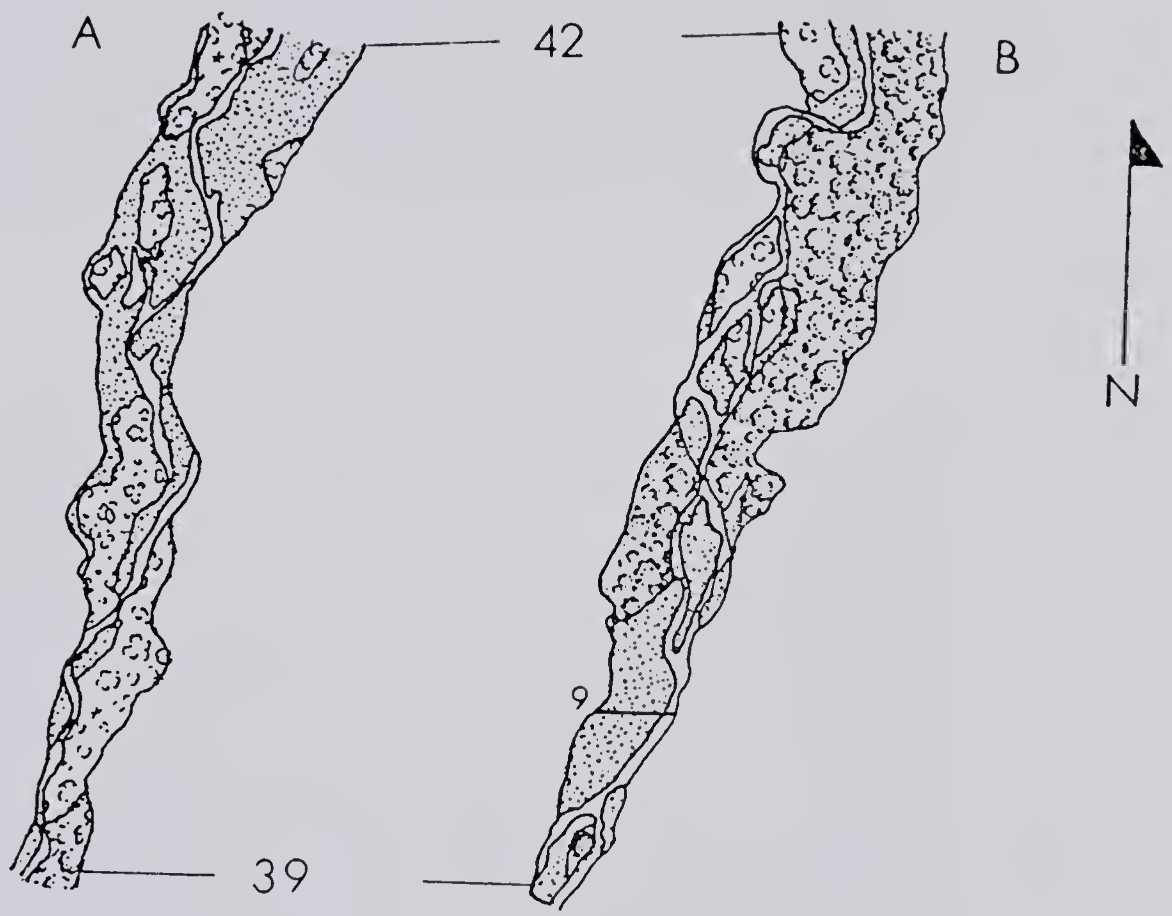
Stands of spruce and some aspen poplar

The first class encompasses the major bar forms mentioned in the text. The second group includes the low surface islands with a transition to the high surface islands which are contained within the last two classes.

Discharge at the time of photography (from the gauging station near Sundre): A = 437 c.f.s. (12.4 c.m.s.)

B = 591 c.f.s. (16.7 c.m.s.)

9 ——— Channel Cross-Section



A
42
B
N
9
39

Figure 5.5 C and D Oblique, aerial photographs of the Red Deer River at sub-unit 4. C looks upstream towards Sundre. D is looking downstream. Note the increase in islands downstream.

C

D

Figure 5.5 E High surface island near km. 42.

Figure 5.5 F Low surface island in foreground with high surface island in the background. Photograph was taken during period of high flow.

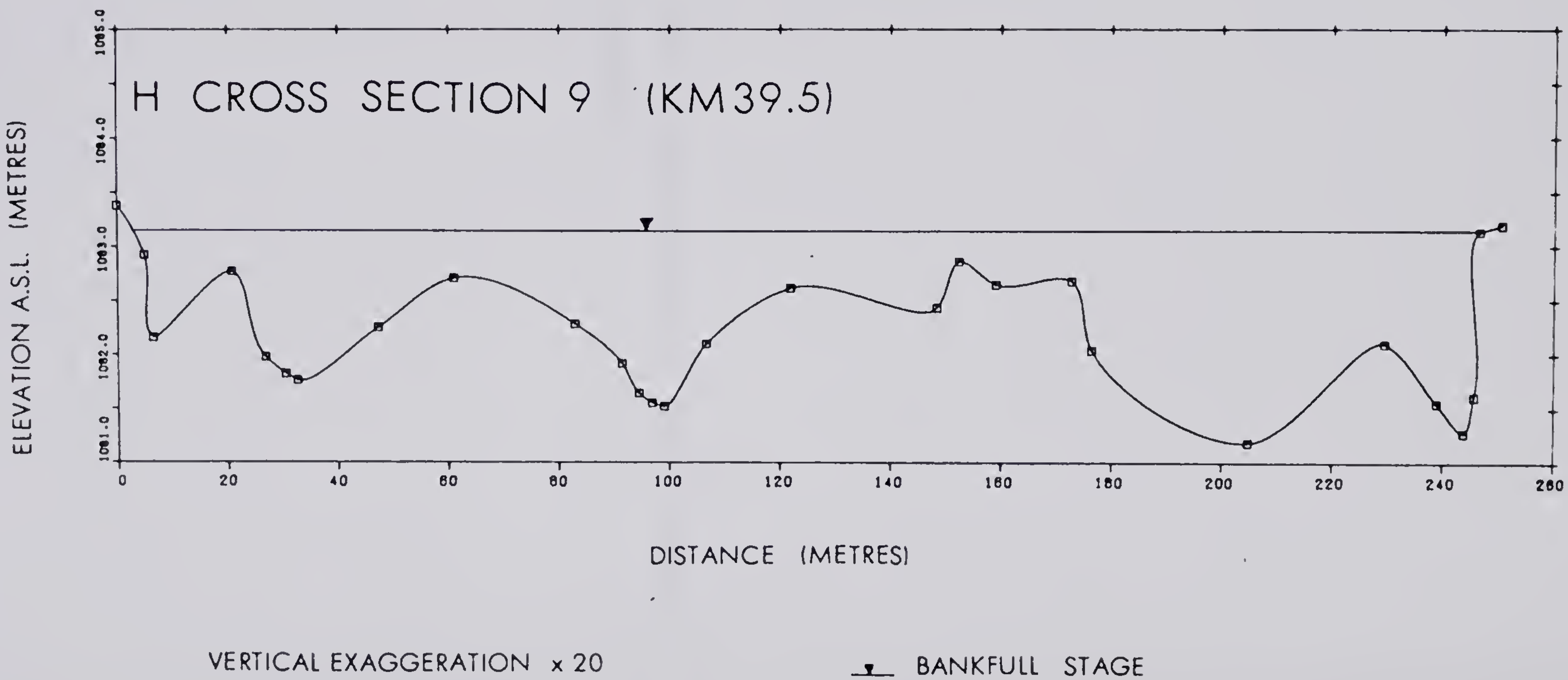


Figure 5.5 G Cross section 9

Figure 5.5 H Cross section 10 (note change of scale).

(see back jacket)

islands have characteristics similar to the floodplain bordering the active channel (Figure 5.6 G). The low surface islands are composed of gravel but lack the layer of finer material characteristic of the high surface form. The vegetation is again sparse, consisting of perennial grasses and isolated stands of willow and aspen poplar (Figure 5.6 D and F).

The bar forms include mainly attached point bars and midchannel longitudinal types. The point bars range in development from embryo types, such as those shown in Figure 5.6 E, to well developed forms (Figure 5.6, C, G and H). These point bars are not characteristic of the "classical" point bar type, with the ridge and depression topography, but rather have the same morphology as those previously described. The midchannel bars are usually found in conjunction with the straight and slightly sinuous channels (Figure 5.6 I and J). Relatively few occurrences of the midchannel braid bars were found. Figure 5.6, K and L, illustrate two channel cross sections from this sub-reach. Figure 5.6, K, cross section 11, is located in an area bordered on one side by bedrock, while cross section 12 (Figure 5.6, L) is contained within gravel banks. Cross section 11 has a single, active channel with a clearly defined thalweg, while cross section 12 has numerous bars and islands, located at various heights relative to the bankfull stage. Numerous well defined channels at various elevations may also be seen in the latter channel cross section.

Channel migration activity has varied throughout this sub-reach. From km 45 to km 55 the rates have been low and have led to relatively minor amounts of bank erosion. Some point bar development has occurred

Figure 5.6 Floodplain modification in sub-unit 5.
Please see back jacket.

A is an interpretation based on aerial photographs from 1963 (scale approximately 1:30,000).

B is an interpretation based on aerial photographs from 1972 (scale approximately 1:21,000).

Scale of A and B is approximately 1:3,000.

LEGEND

Exposed gravel

Perennial grasses with scattered stands of willow and immature aspen poplar

Dense stands of aspen poplar

Stands of spruce and some aspen poplar

The first class encompasses the major bar forms mentioned in the text. The second group includes the low surface islands with a transition to the high surface islands which are contained within the last two classes.

Discharge at the time of photography (from the gauging station near Sundre): A = 437 c.f.s. (12.4 c.m.s.)

B = 591 c.f.s. (16.7 c.m.s.)

11 —— Channel Cross-Section

Figure 5.6 C,D and E Various channel patterns and vegetative patterns found in this sub-unit.

C

D

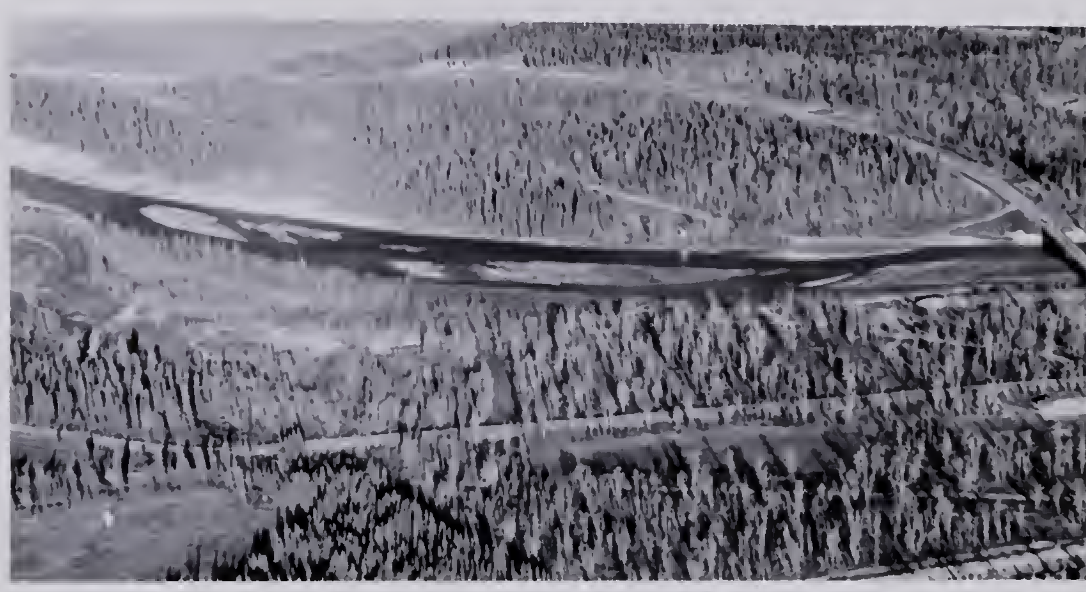
E

Figure 5.6 F A dissected point bar. Note the increase in the vegetation development towards the bank.

Figure 5.6 G A large high surface island near km. 75. Note the secondary channels across the surface.

Figure 5.6 H,I and J Various bar forms found in the sub-unit.

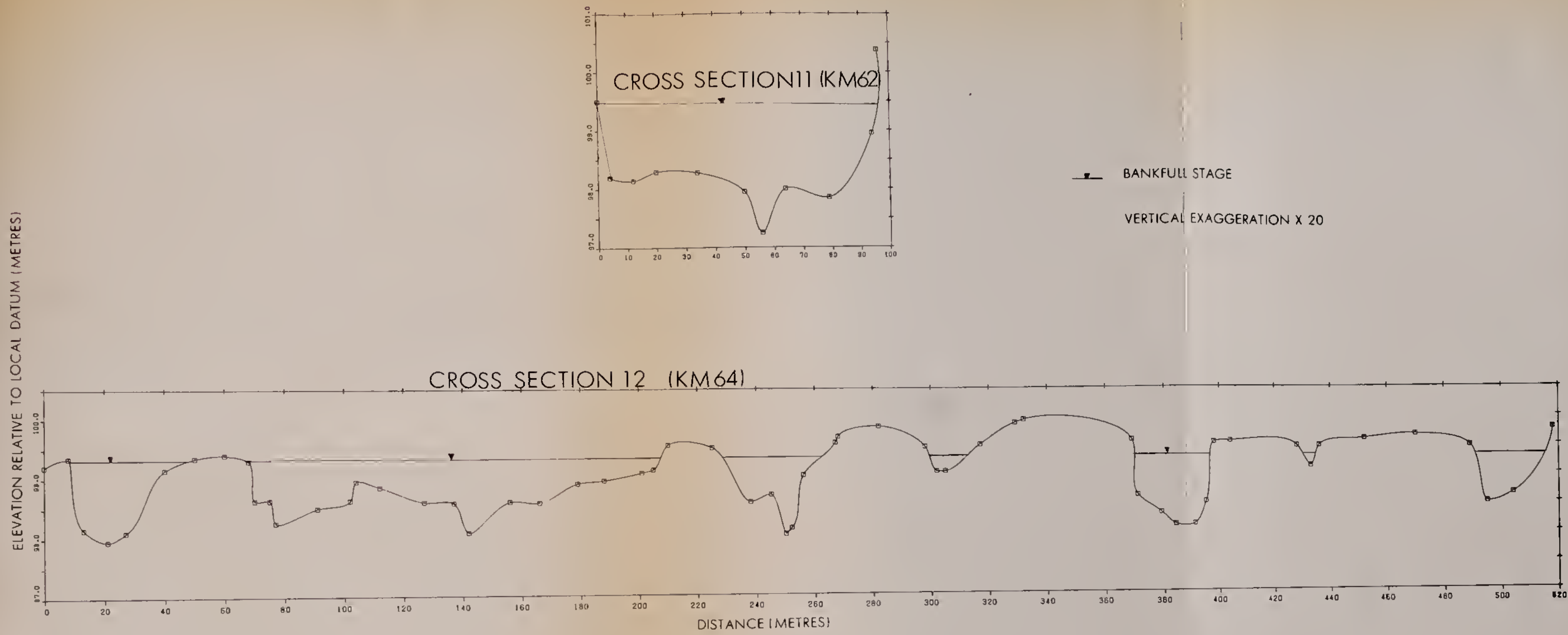
H



I

J

Figure 5.6 K and L Cross sections 11 and 12.


CROSS SECTION11 (KM62)
BANKFULL STAGE
VERTICAL EXAGGERATION X 20
CROSS SECTION 12 (KM64)
ELEVATION RELATIVE TO LOCAL DATUM (METRES)
DISTANCE (METRES)
97.0
98.0
99.0
100.0
101.0
0
10
20
30
40
50
60
70
80
90
100
120
140
160
180
200
220
240
260
280
300
320
340
360
380
400
420
440
460
480
500
520

during the time separating the aerial photographs and from km 55 to km 61 an increase in the rate of channel migration has taken place. This has resulted in the erosion of large areas of the floodplain. The small reach from km 61 to km 63 has experienced little erosion and channel migration. The floodplain from km 63 to km 72 shows a moderate amount of channel migration over a ten year period. Channels have incised into a number of point bars creating islands. The final subsection, from km 72 to km 75, reflects little evidence of change over a ten year period. Some active point bar formation has occurred although this does not appear to have had a significant effect on the channel displacement.

5.2.3 DISCUSSION

Some general conclusions may be drawn from the descriptions of the sub-reach islands in the upper Red Deer River. The high surface type is similar to the floodplain in terms of both sedimentary and vegetation characteristics; these islands are in effect isolated remnants of the floodplain bordering the active channels. Reactivation of formerly abandoned channels, or the formation of new ones, have resulted in this separation. Islands of the low surface variety represent remnants of point bars after their dissection. This is evident on the maps derived from the sequential aerial photographs. These photographs show that reactivation of the former channels, incised across point bar surfaces, results in the progressive abandonment of the main courses in favour of the newly activated ones. The low surface

islands are relatively transient in nature, their relative degree of stability being intermediate to the extremes of the active bars and the high surface islands.

A comparison of the bar and island descriptions, with those of the channel patterns, shows that the high island forms are associated with the wider portions of the study reach. This is especially true in Zones 2, 4, and 5 (Figure 5.1). Zone 1 has a number of high island forms but these are generally restricted to the area between km 5 and km 10, which has a higher width than the adjacent sub-reaches. Zone 1 is at present being incised into bedrock and does not display a well developed floodplain throughout much of its length. Zone 3, which extends from km 30 to km 42, experiences some of the high surface island forms, but these are restricted to the downstream sector of the sub-reach.

The most prominent bar form found along the study reach is the point bar, with the exception of Zone 1 where it occurs in only a few restricted areas. The best developed forms are associated with portions of the active channel which have moderate Wr values and occur in association with the bedrock outcrops. The low sinuosity segment, from km 30 to km 39, shows point bars in their initial stages of development. Midchannel bars are also found throughout the study area. The braid bars are most restricted in their development, occurring only where active braiding is relatively intense. The longitudinal bar, on the other hand, does not appear to be restricted to any one channel pattern or area.

Areas dominated by channel migration through bank erosion are mainly confined to single channel, sinuous reaches. Channel migration through avulsion is the predominant process found in the braided reaches. In such areas, previously abandoned channels are periodically reoccupied and reactivated. Major shifting of the channels is achieved at the expense of other active ones.

Sectors of relatively intense channel migration tend to be associated with zones of high and variable Wr values - where braiding is pronounced. This is especially true of Zones 2 and 4. The migration rates in Zone 2 appear to have been relatively high as is the intensity of braiding. In Zones 4 and 5 the degree of channel migration varies markedly. The areas of most intense channel migration are all found upstream from the bedrock outcrops, an important characteristic which will be discussed more fully in the final chapter.

5.3 LONGITUDINAL PROFILE

5.3.1 INTRODUCTION

In his discussion of a graded stream, Mackin (1948, p. 471) stated that:

> ...over a period of years, slope is delicately adjusted to provide, with available energy and prevailing channel characteristics, just the velocity required for the transportation of the load supplied...The graded stream is a system in equilibrium...

He goes on to say that changes in the variables controlling the longitudinal profile, discharge and sediment supply, will produce a shift

in this equilibrium in the direction which will minimise or eliminate the effects of the perturbations. For example, an increase in the sediment supply will promote localised steepening of the channel slope by enhancing deposition. Aggradation will continue to increase the slope until velocities competent to transport the extra load are reached. Mackin's (1948) use of the term "over a period of years", as pointed out by Dury (1966), eliminated the need for considering shorter term fluctuations. These short term, daily to seasonal, fluctuations were considered by Rubey (1952) to be of substantial importance and he noted that a river is never "graded" because of these fluctuations. Equilibrium is primarily determined by discharge and the sediment load, as opposed to the distance from the source of the river (Rubey, 1952). Leopold and Maddock (1953) found that a decrease in particle size is indicative of a decrease in the velocity, which is the result of a decrease in the channel slope. They interpreted the longitudinal profile to be determined largely by the physiographic history of the area and that adjustments to variations in discharge and sediment load are accomplished through changes in the channel cross sectional geometry.

Leopold, et. al., (1964) believed that the slope of a river is a function of variables related to the hydraulic geometry of a channel, such as width, depth, and velocity. They reiterated Hack's (1957) conclusion that the local geology has a significant influence on the channel slope and that different lithological properties result in the development of specific longitudinal profile characteristics.

Dury (1966) claimed that occurrences of braided reaches in

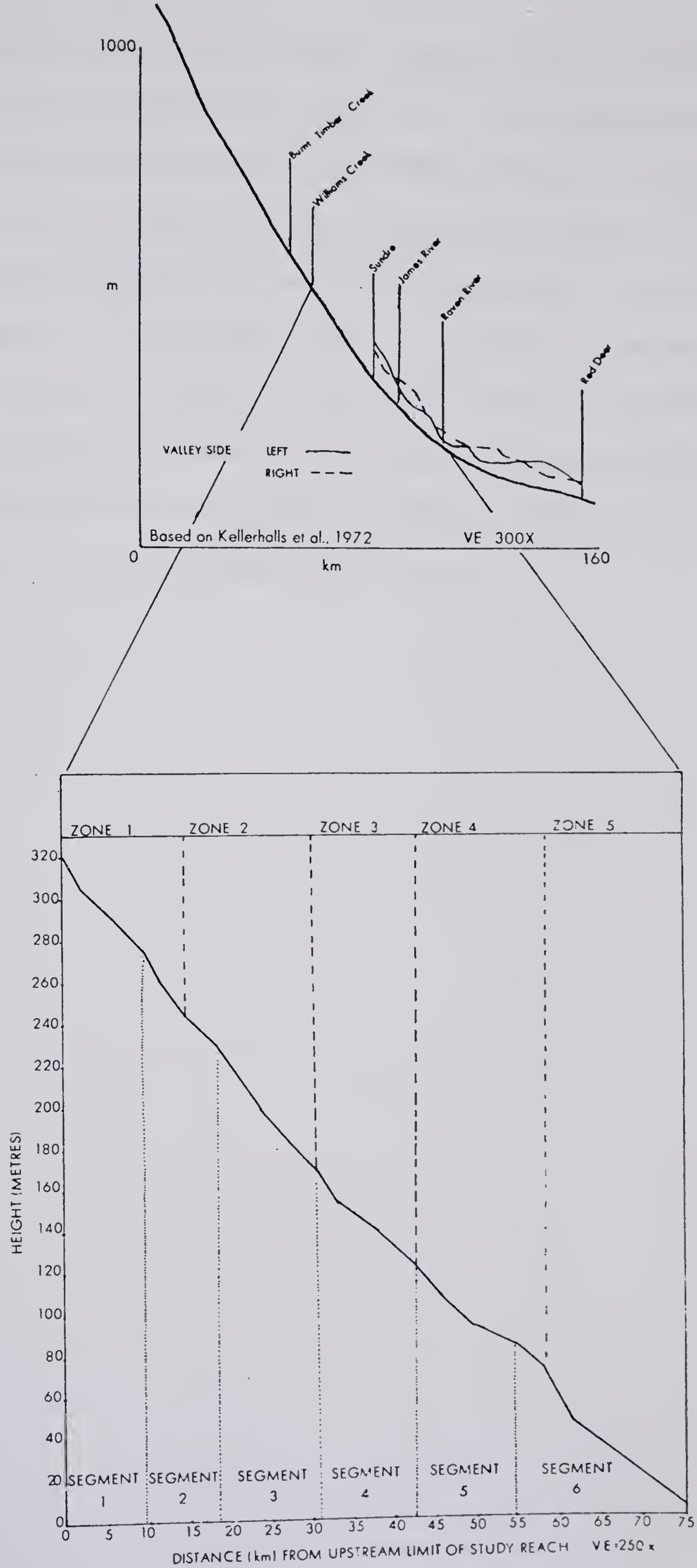
a predominantly meandering river, and meandering reaches in a braided one, are sufficient to cause breaks in the longitudinal profile. He stated that a change in bank materials or a bedrock outcrop could inhibit the development of channel depth, thus favouring width and changing the longitudinal profile significantly. Therefore, a river flowing through a region composed of a variety of bedrock lithologies and bank materials will rarely develop a smooth longitudinal profile, but rather form a series of segments which strongly reflect the local controls of varied bank and bed resistances.

5.3.2 LONGITUDINAL PROFILE OF THE STUDY REACH

The measurements for the derivation of the longitudinal profile of the study reach were made from topographic maps (scale 1:50, 000) because surveying of the channel slope was not possible during the time available for field work. It is realised that topographic maps at this scale are not ideally suited for detailed longitudinal profile analysis but they do yield the general trends shown in Figure 5.7. This figure was deliberately drawn with a large vertical exaggeration so that minor irregularities of average sub-reach slopes might be accentuated.

As for most rivers the longitudinal profile of the entire study reach is concave-up, although this concavity is not particularly well defined. Of greater significance are numerous breaks in channel slope along the profile which demarcate a series of smaller concave segments, designated 1 to 6 in Figure 5.7. Segments 1 and 2 are located mainly in the foothills area. Their relatively steep slopes

Figure 5.7 Longitudinal profile of the study reach.


1000
m
Burnt Timber Creek
Williams Creek
Sundre
James River
Raven River
Red Deer
VALLEY SIDE
LEFT
RIGHT
Based on Kellerhalls et al., 1972
VE 300X
0
km
160
ZONE 1
ZONE 2
ZONE 3
ZONE 4
ZONE 5
320
300
280
260
240
220
200
180
160
140
120
100
80
60
40
20
0
HEIGHT (METRES)
SEGMENT 1
SEGMENT 2
SEGMENT 3
SEGMENT 4
SEGMENT 5
SEGMENT 6
0
5
10
15
20
25
30
35
40
45
50
55
60
65
70
75
DISTANCE (km) FROM UPSTREAM LIMIT OF STUDY REACH
VE=250 x

occur in conjunction with low Wr values. Several smaller subsections of the profile may be identified. Among these is one beginning at km 2 and extending downriver with a lower-angle gradient to km 9.5. In this section the Wr values increase substantially (Figure 4.1). A similar association applies to the section of Segment 2 which begins at km 15. Segment 3, from km 18 to km 30, incorporates a number of these smaller sections. It will be noted from Figure 4.1, however, that a major change in the channel pattern does not occur within this segment. A return to the basic pattern, noted for Segments 1 and 2, is found in the final three segments, 4, 5 and 6.

CHAPTER SIX

DISCUSSION AND CONCLUSIONS

6.1 DISCUSSION

If the factors outlined in the previous chapters are considered together the division of the study reach into separate zones, based on channel characteristics as outlined in Chapter Five, (Figure 5.1), may be discussed in greater detail. Figure 6.1 illustrates this zonation with summarised information on the respective Wr values, channel slopes and bank and channel material characteristics. There appears to be a general trend towards the expected, with higher Wr values, lower average slopes and finer channel materials occurring downstream. The Wr values are partly deceiving, though, as is evident by comparison of Figures 5.1 and 6.1. Zones 2, 4, and 5 were interpreted earlier as being truly braided over most of their areas. As indicated in Chapter Five, Section 5.3, there is little correlation between the numerous breaks in the channel slope and changes in the channel patterns. Zone 3, Figure 6.1, has a moderate slope but does not display the high and variable Wr values that the adjacent zones do. The first, Zone 1, is incised mainly into bedrock and has a generally low Wr value.

The bar and island classification of Chapter Five distinguished low and high surface islands. Comparisons of the distribution of these forms shows that the high surface forms tend to be concentrated in

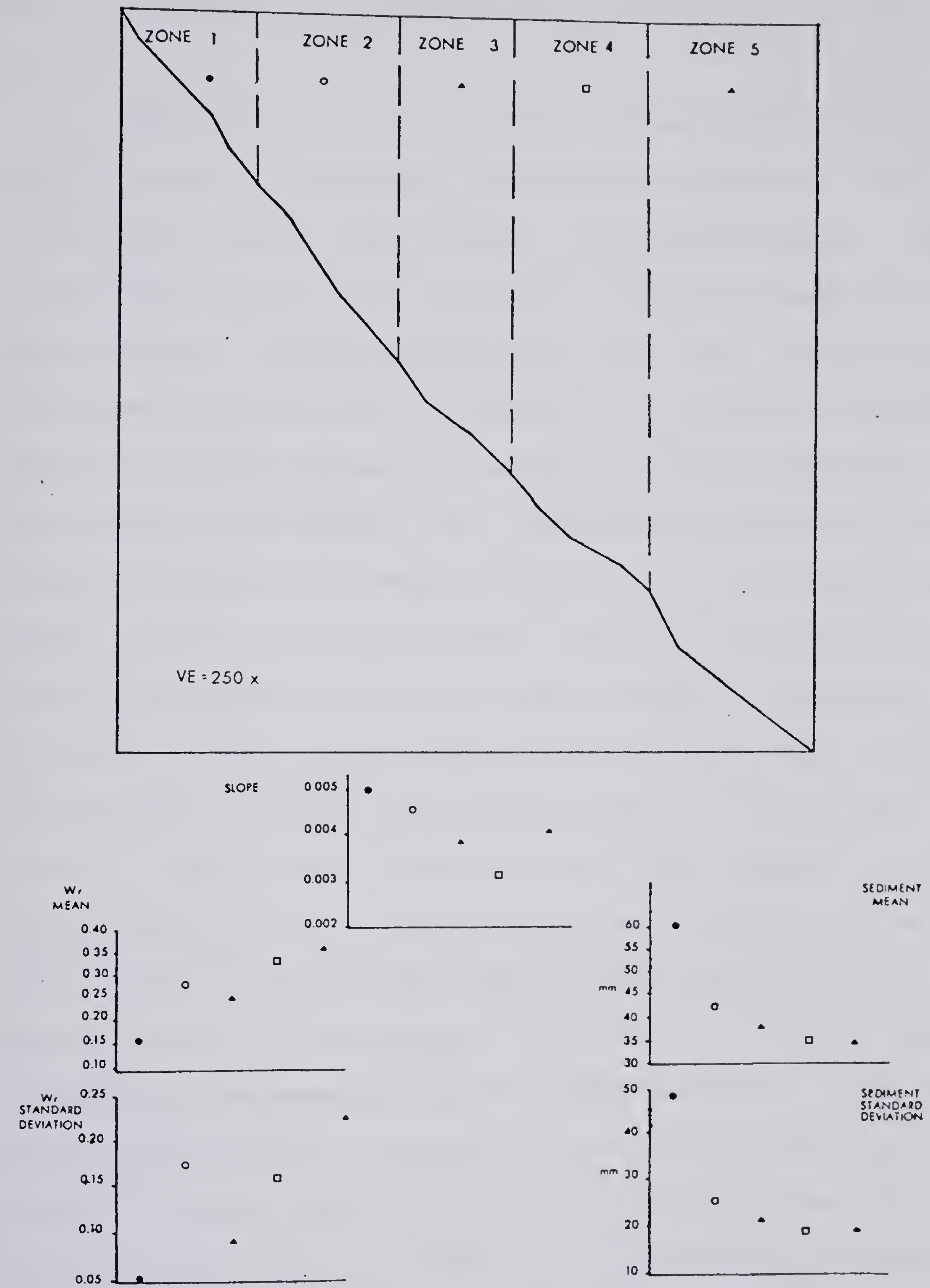


Figure 6.1 Composite diagram based on Wr divisions.

the lower portions of the sub-reaches but are infrequent in Zones 2 and 3.

Figure 6.1 does not fully reflect the variations in the channel patterns observed. The numerous breaks in the longitudinal profile do not closely relate to major changes in the channel pattern. Figure 6.2 delineates the secondary long profile concavities discussed in Chapter Five,with summarised information on Wr values, average slope, and sediment characteristics. In addition the locations of prominent bedrock outcrops are indicated. Interesting patterns arising from this diagram are as follows. First the lower slope segments of the secondary concavities have higher Wr values, as well as greater variability, than do the steeper segments. Second, the channel material size and sorting vary within these segments as well. In Segments 1, 2 and 5, as well as part of Segment 3, the steeper channel slopes are associated with larger mean sediment sizes than for the lower slope portions. Third, the major bedrock outcrops usually delimit the transitions between the main secondary concavities. Upstream from each bedrock constriction the channel slope is comparatively low, while downstream there is a steepening of the gradient. The lower - angle slope segments are associated with the higher Wr values while at the bedrock constrictions, and immediately downstream from them, the values decrease. Numerous islands tend to be concentrated upstream from the bedrock outcrops. The areas immediately upstream from the outcrops, therefore, appear to be sites of accentuated aggradation. The effect of the bedrock, in most cases, is to divert the direction of flow of the river. During high flows the bedrock constrictions assume their

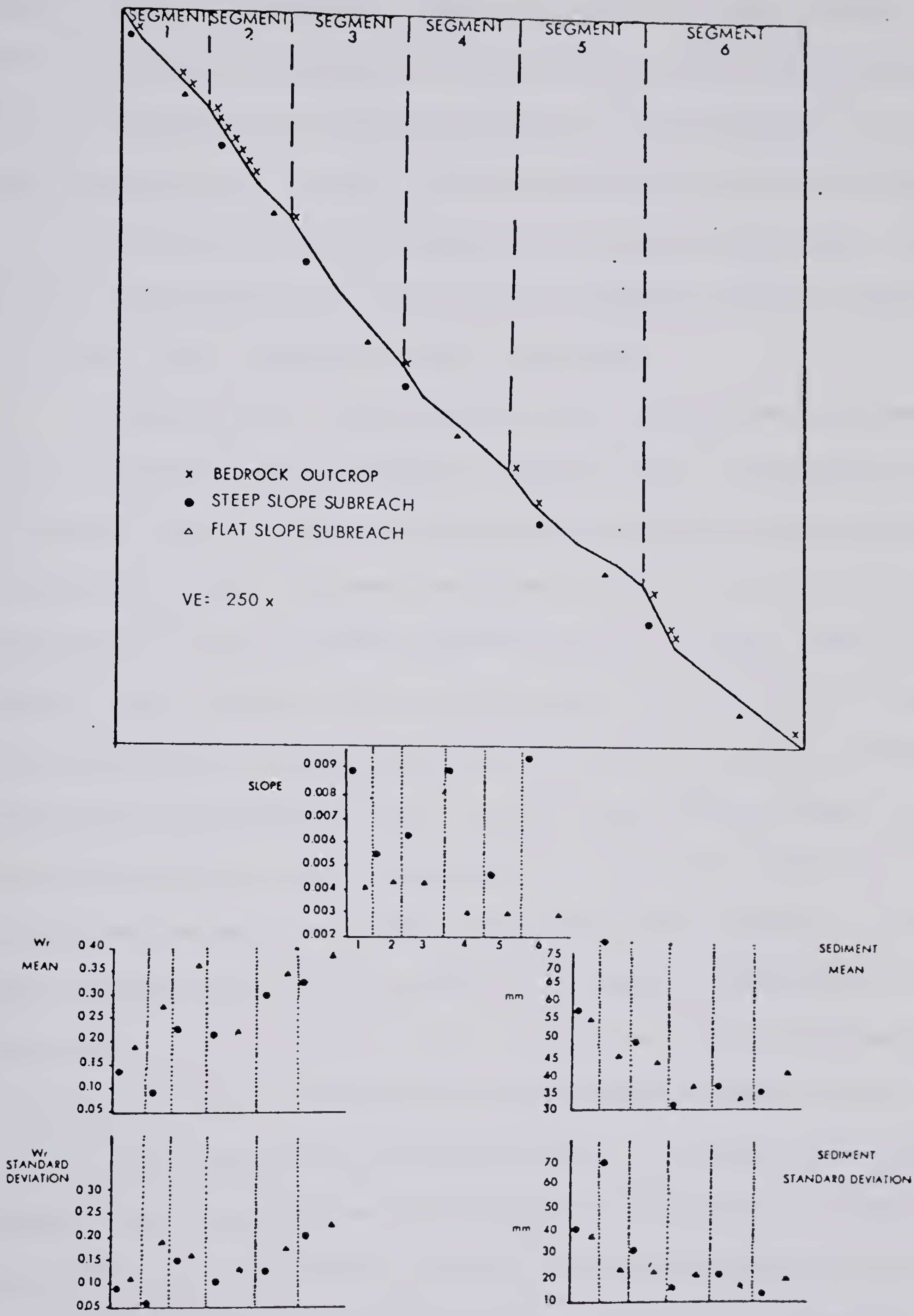


Figure 6.2 Composite diagram based on secondary concavities in longitudinal profile.

greatest relative importance, promoting partial backwater effects. This results in the preferential development of high surfaces islands by the reactivation of abandoned channels on the floodplain. The assumed decrease in local flow velocities immediately upstream of the bedrock constructions probably enhances local sediment deposition. This leads to the comparatively rapid channel migration, and bar and island evolution, which characterise such sub-reaches.

The accentuated aggradation is also evident from the channel cross sections presented earlier in Chapter Five. The channels located in the sub-reaches with higher-angle slopes tend to be substantially narrower and deeper (see cross sections 2, 3, 4, 5, 7, 8 and 11) than those found in the low-angle slope portions of the study reach. Wide channel cross sections (see cross sections 1, 6, 9, 10 and 12) show numerous channels separated by bars and islands at various elevations relative to the bankfull level. Not only are the mean channel slopes relatively gentle in these aggradational sections but the relative channel depths decrease. This relationship should increase the relative roughness and flow resistance of the channels (Church and Gilbert, 1975). In turn these tend to inhibit the flow velocities, promoting deposition of the material transported into these areas.

Downstream from the bedrock outcrops increases of the mean channel slope may stimulate the relative effectiveness of sediment transport. These increases in channel slope are accompanied by decreases in width and increases in depth. This combination is most likely to accentuate sediment transport efficiency. The sediment size characteristics of each segment, illustrated in Figure 6.2, may

be explained in this context. First, the inferred increase of the relative transport efficiency through the steeply sloping sub-reaches results in the delivery of finer materials downstream. Second, the limited local supply of finer gravel, in sub-reaches bordered by bedrock, ensures that there will be a dominance of coarser clasts. The winnowing of the available finer gravel eventually results in the development of a coarse-clast armoured bed while finer materials are deposited further downriver in the zones of lower channel slopes. This is best displayed by segments 2 and 3, Figure 6.2. In the upper sections of these sub-reaches there is a slight downriver decrease in the mean sediment size along with a decrease in the slope. The third section, of segment 3, shows a decrease in the size of the channel materials with an increase in the slope. The average Wr value also decreases significantly. The reason for this deviation is evident from Figure 4.1. At this location width development is restricted by the occurrence of relatively cohesive lacustrine material. This restriction in width is associated with an increase of the channel slope. Segment 6 follows the pattern developed for the previously discussed sections in terms of slope and width. However, there is a deviation in terms of sediment size gradation. This segment has a finer-grained sediment component in the upper, higher slope segment than is found in the lower sections. The same is true for Segment 4. These two segments occur directly downstream from the confluence of two main tributaries - Fallentimber Creek in the case of Segment 4, and James River for Segment 6. The substantially finer materials in these steeper sub-reaches probably reflect abnormally large deliveries

of sediment to the Red Deer River by the two major tributaries,

Thus some general conclusions are possible, First, there appears to be a subtle correlation between channel slope and channel pattern, as was pointed out in Chapter Five and earlier in this chapter, This relationship is weak, however, because of a number of extraneous factors such as the varied nature of bank materials within the study reach. Second, the occurrence of bedrock outcrops appears to strongly dictate channel pattern development, The location of major bedrock outcrops at the upstream and downstream limits of the secondary concavities is more than coincidental, Dury (1966) suggested that such breaks in slope may also be an effect, rather than a cause of river behaviour. He stated that a change in channel pattern from one which is predominantly braided to sinuous, and vice versa, is sufficient to affect marked discontinuities of the longitudinal profile. Evidence suggests that for the upper Red Deer River the main factor contributing to the breaks in the long profile is the distribution of the bank materials. Changes in the channel slope are not closely related to major variations of the channel pattern, This is illustrated, for example, by Segment 4, sections 1 and 2, (Figure 6.2) where no discernable change in the Wr ratios accompanies the break in slope. It is concluded that although the channel slope may be partly instrumental in determining the channel pattern other factors such as the nature and distribution of the various bank materials have equal, if not greater, importance in the case of the upper Red Deer River.

Another generalisation concerns the variations of sediments contributed to the river, For example, Segment 2 straddles the east-

ern boundary of the foothills region and the western portions of the Bearberry Prairie. As shown by Figures 4.12 to 4.15, a distinct break in the trend of the sediment size occurs at this point (km 22). This is also clearly noted in Figure 6.2. Other segments within geomorphologically distinct areas do not show this obvious difference. In these areas, although variations do exist in the slope-sediment size relationship, the variations are not great. These areas have sources of sediment which contribute material comparable in size to each section, so that large differences are not expected. The rather large average size of sediment found in the lower portions of Segment 6 may be explained by contributions from a different geomorphological source, that is the outwash fan/delta described in Chapter Three. This will have had the effect of adding material not previously reworked by the Red Deer River to the degree found in the other sections.

The poor evidence of channel migration activity in Zones 1 and 5 (Figure 6.1) suggests that the maximum sediment sizes in these zones exceed the local river competence. This has led to the development of lag pavements along the sides of the islands, preventing further erosion of the island banks except during exceptionally high flow events. This further emphasises the point regarding the introduction of coarse material into the river in the downstream parts of the study reach. In other areas channel migration has been relatively intense, particularly in the sectors of accentuated aggradation where finer gravels are more prevalent.

An interpretation of the sequential aerial photography, from 1957-58, 1962-63, and 1972, indicated that the channel migration pat-

terns have varied considerably through time. Little difference was evident for the patterns of the 1957-58 and the 1962-63 photographs but substantial changes were found between the 1962-63 and 1972 photographs. An examination of the discharge records revealed the main reason for this disparity in channel migration activity. Table 6.1 shows the maximum instantaneous discharges at Sundre for the 1957 to 1972 period. It was not until 1965, during this period, that the estimated bankfull discharge of 12,000 c.f.s. (340 c.m.s.) was attained or exceeded. Additional, major floods occurred in 1967 and 1970. The records for the years 1957 to 1965 indicate that the peak discharges were well below the bankfull level. It appears from this that major formative discharges must be close to or greater than bankfull discharges in the case of the upper Red Deer River. Channel migration and major bedload sediment transport events are governed by the bankfull or near bankfull discharges. This is in agreement with the conclusions presented by Harvey (1969) who found that the streams with an important peak flow component had the channel capacity and geometry adjusted to their bankfull discharges.

The highly variable discharges typical of the study reach engender marked variations in the transportation of bedload material. These variations involve not only the amount of material transported but also the calibre of the load. Evidence of this lies in the occurrence of distinctive sheets and lobes of well sorted gravels. Figure 6.3 shows an example of sheet deposits on a point bar surface. Figure 6.4 A and B, illustrate lobate phenomena within secondary channels. Somewhat similar gravel sheets and lobes have also been reported

Table 6.1 Maximum instantaneous discharges for the Red Deer River at Sundre.

YEAR	Discharge C.F.S.	C.M.S.
1957	2,960	82.9
1958	4,030	113.4
1959	8,010	224.3
1960	3,390	94.9
1961	4,990	139.7
1962	2,730	76.4
1963	6,250	175.0
1964	9,780	273.8
1965	23,100	646.8
1966	8,580	240.2
1967	16,400	459.2
1968	-	-
1969	10,000	280.0
1970	18,000	504.0
1971	7,850	219.8
1972	10,200	285.6

(Source: Canada Department of Fisheries and Environment, 1977, p. 269).

Figure 6.3 Diffuse gravel sheets on point bar surface.

A

B

Figure 6.4 A and B Gravel lobes within secondary channels.

by Gustavson (1978) from the Nueces River, Texas. These forms appear to be the product of the passing of a floodwave, where bedload materials are deposited during the declining stages of the flood. The size distribution of the sediments within the lobes in particular should reveal systematic vertical and downriver variations if this deduction is correct. The vertical changes in the grain sizes should reflect a progressive decrease in the flow velocities after the passing of the flood crest, while the downriver variation should reflect finer material being transported further than the coarser sediment.

The gravel lobes were located close to the confluence of the smaller secondary channels and the main channel. The greatest thicknesses of the lobes were apparently close to the junction of the two channels, becoming thinner upstream. This possibly implies that the lobes are temporary reservoirs for the smaller gravel sizes. During the moderate discharge events water flowing faster through the main channel may develop a partial backwater effect in the tributary channel, causing deposition of the gravels within it. At higher discharges the velocities within the smaller channels may become sufficiently high to transfer part or all of the lobe material into the larger primary channel. When this occurs the sediment becomes less concentrated and may be eventually deposited in the form of thin, diffuse sheets of gravel on point bar surfaces further downriver.

6.2 CONCLUSIONS

General conclusions regarding the development of channel patterns

in the upper Red Deer River Valley are as follows.

(1) The bank materials, although mainly comprised of gravel also embody numerous bedrock outcrops and a minor amount of lacustrine silt and clay. These materials and their distribution are closely related to braided channel pattern development. The location of bedrock and lacustrine materials at the boundary of the secondary concave segments is more than coincidental. Where gravels are absent braiding is very poorly developed.

(2) Braided channels tend to accompany relatively gentle channel slopes. These sectors are usually located immediately upstream from major bedrock constrictions. The degree to which these constrictions have influenced the development of the more gently-sloping segments is not known with certainty. It is apparent that the breaks in the secondary concavities are not caused exclusively by changes in the channel pattern. As was noted for Segment 1, for example, the change in the channel pattern occurs downstream of the break in the slope. The areas of gentler gradient may partly reflect unknown, long term elements of the river valley's evolution but the most obvious contemporary control is that of the bedrock constrictions. Further emphasis of these areas is achieved by contemporary, enhanced deposition of material transported from the steeper slope sections.

(3) The discharge patterns typical of this region include numerous

high flow events during the early summer period. The magnitude and duration of these events vary markedly from year to year, depending on variations of the climatological conditions. Such variations determine episodes of relatively intense channel migration, the periods of highest flows are also the times when bedload transport dominates. It is believed that only during the times when bankfull discharges are approached, or exceeded, is bedload transport and deposition on a scale sufficient to markedly influence the channel migration.

(4) The islands found in this study reach are of two main types. The high surface islands are the result of the separation of portions of the floodplain by the reactivation of formerly abandoned channels. Those classified as low surface forms are a result of incision and erosion of point bars, and other gravel surfaces, located within the active channels. These forms are concentrated primarily in areas upstream from the bedrock controls, that is, in areas which are braided.

It is apparent from this study that much remains unknown about the behaviour of the upper Red Deer River. Some of the more important problems relate to the hydraulic factors characterising different sub-reaches, and the sediment transport rates within these. An additional problem is the role of seepage and its effects on the hydraulic regime of the river, especially in the vicinity of the Bearberry Prairie. These problems were beyond the scope of this study but may form the bases of further research.

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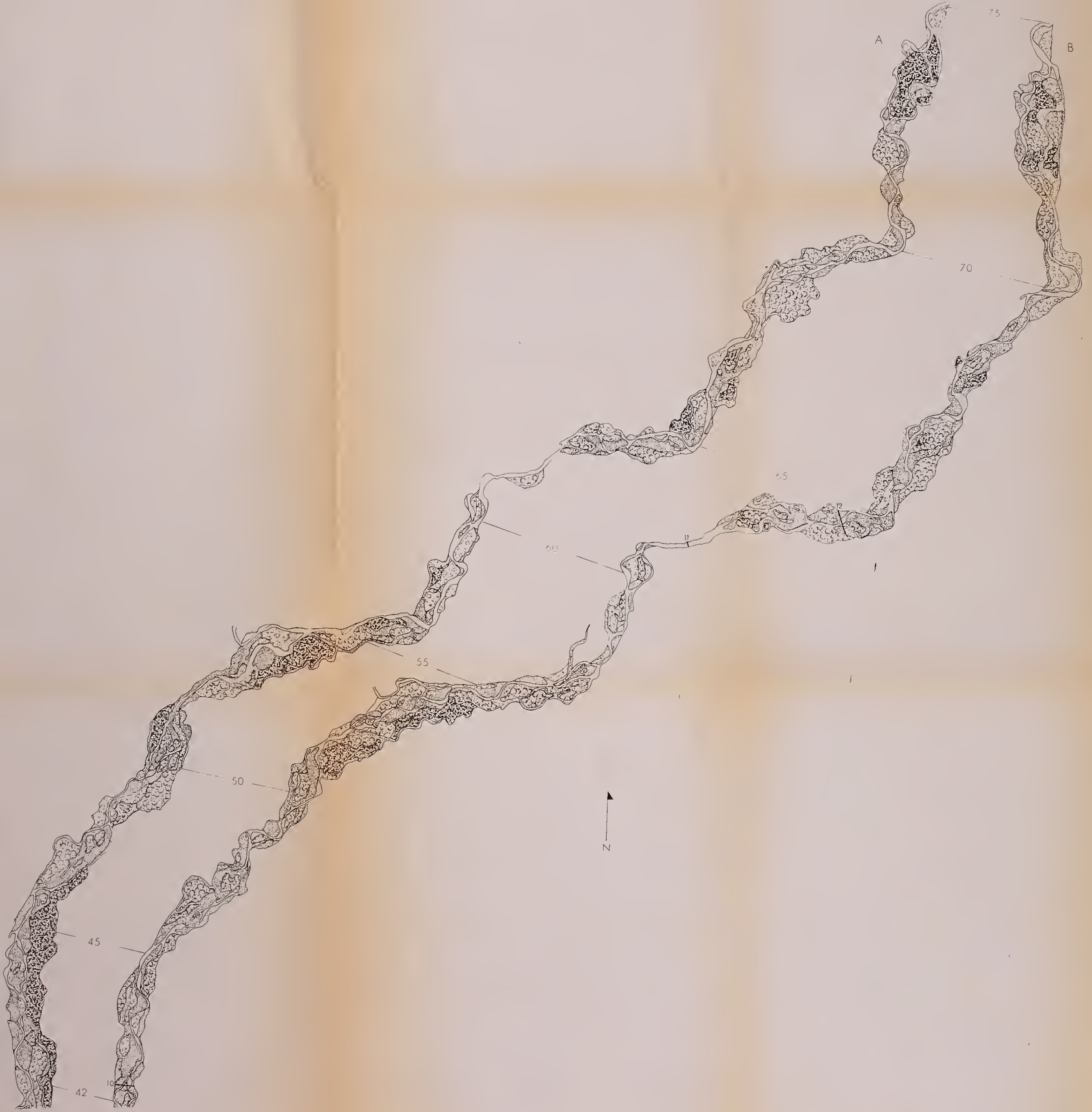
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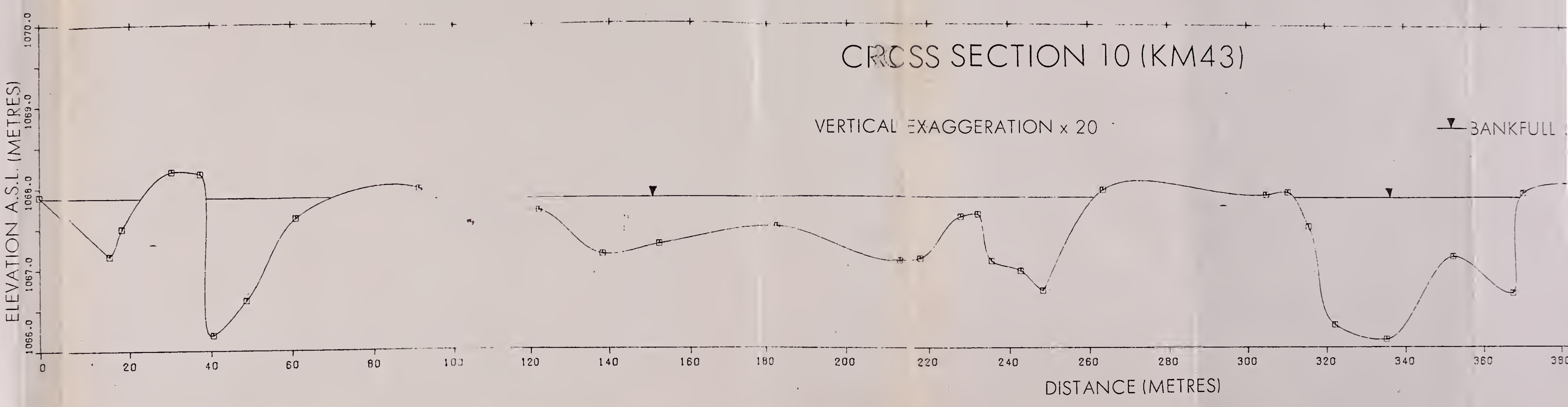
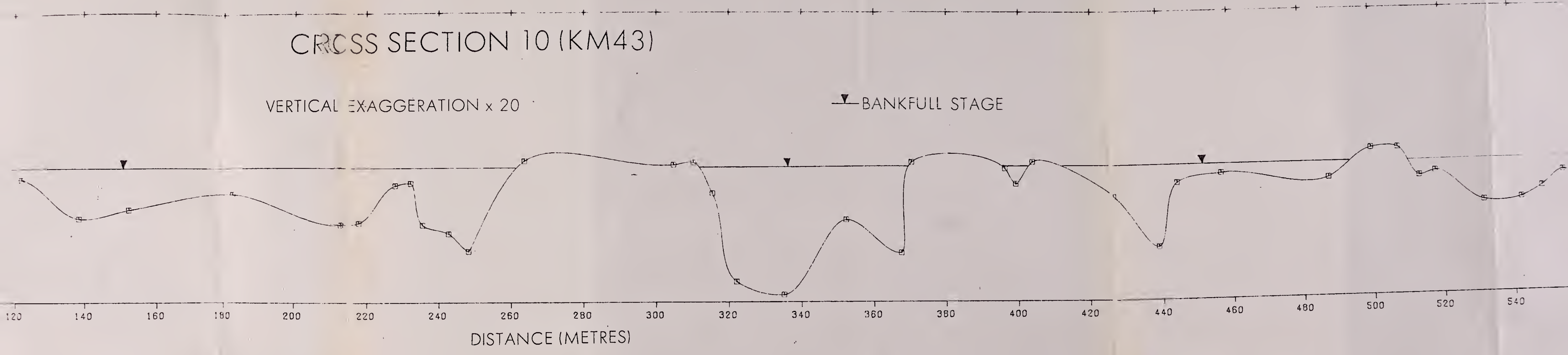
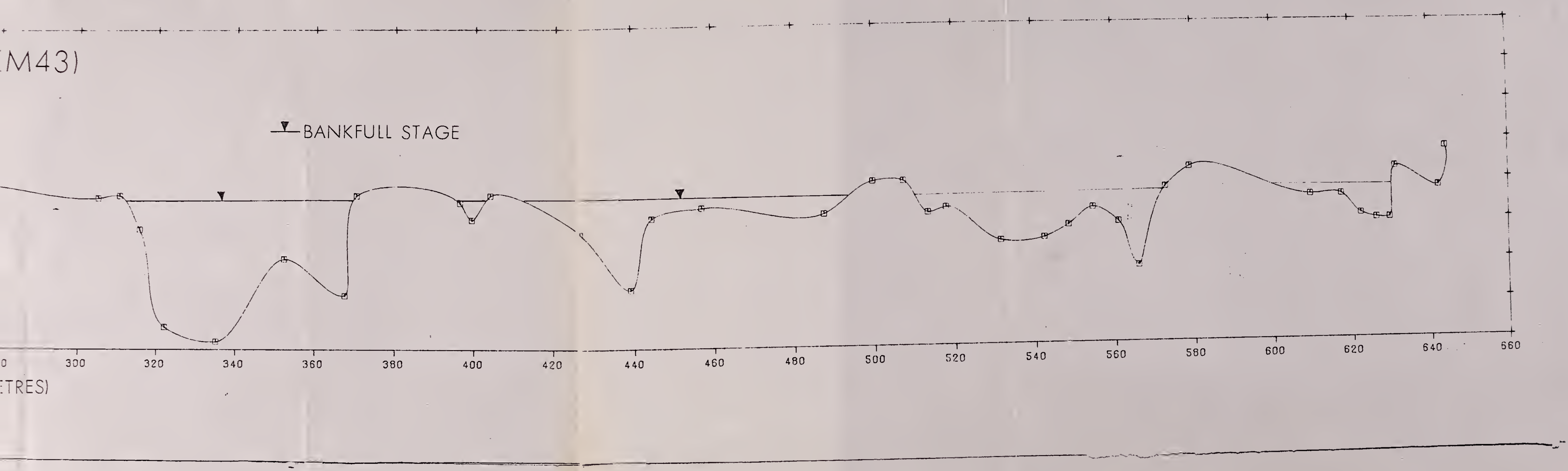


CROSS SECTION 10 (KM43)
VERTICAL EXAGGERATION x 20
BANKFULL
ELEVATION A.S.L. (METRES)
1070.0
1069.0
1068.0
1067.0
1066.0
0
20
40
60
80
100
120
140
160
180
200
220
240
260
280
300
320
340
360
380
DISTANCE (METRES)

CROSS SECTION 10 (KM43)





M43)
BANKFULL STAGE
300
320
340
360
380
400
420
440
460
480
500
520
540
560
580
600
620
640
660
ETRES)